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THE GENETIC SIGNIFICANCE OF SOME TIN–TUNGSTEN LODES IN STEWART ISLAND, NEW ZEALAND.

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GEOLOGIC SETTING.

STEWART Island is a rugged, bush-clad island lying off the south coast of New Zealand. It is composed mainly of granitic rocks forming a section of a large granite batholith that appears on

the neighboring coast of New Zealand. In the small area described in this paper, the granite is intrusive into a biotite-muscovite-schist with some interlaminated quartzite, but these rocks

FIG. 1. Locality map.

remain only as small roof-pendants. A tin-tungsten lode-zone occurs within one of these pendants; it has no commercial interest, but presents many interesting metallogenetic features.

The granite meets the schist along sharply defined unreactive contacts. Consequently it must be inferred that the batholith penetrated to a comparatively shallow depth: *i.e.*, it rose within an orogeny into the zone of directed pressure.

The granite is predominantly a white biotite-granite (with about 77 per cent. SiO_2), in which plagioclase is subordinate to orthoclase. Muscovite is not a primary mineral. The biotite-muscovite-schist (Fig. 4-D) consists of alternating laminæ of light and dark mica. Quartz is not present in all samples of the schist: it may have migrated by endogenous secretion to the quartzite bands, thus accentuating the contrast between the original psammitic and pelitic beds.

Several types of deuteric reaction were observed in the granite. Almost all thin-sections showed incipient kaolinization and sericitization. Such processes were not unconnected with the introduction of a quartz-magnetite-sericite final ichor that penetrated fracture-systems in the consolidated granite. These late solutions may be regarded as typically hydrothermal and were probably responsible for the introduction of small quantities of gold into the granite, and for the formation of small lead, copper, and manganese veins.

In some localities near the contacts, a more vigorous reconstitution, which may be regarded as a mild greisenization, was effected by higher temperature emanations bearing some fluorine (but no tin or tungsten). The product of this process is a muscovite-myrmekite-granite. This rock occurs sporadically along the roof of the batholith, but it is distinct from the greisen which was injected as a separate petrological entity into the schistose roof, and there differentiated to produce tin-tungsten lodes. The former greisen is the product of (pneumatolytic?) reaction; the latter is the product of advanced differentiation. The mode of intrusion of this final greisenous extract along a zone of weakness in the schist, and its differentiation within that zone, form the subject of this paper.

Structural Position of Lode-Zone.

The position of the metalliferous lode-zone is indicated in Fig. 2. The regional strike of the foliation of the metamorphic rocks throughout the district is N. 70° W., but the metalliferous ichors patronized a belt along the side of one of the schist pendants where a local superposed structure had been impressed (at right angles to the regional structure). The regional structure dips gently southwards, but the superposed structure consists of a series of closely packed isoclinal folds, and may have been impressed by pressure at the forefront of the granite mass now forming the Robertson River basin. Peripheral tension induced about the borders of this mass when it consolidated, formed a tension-zone that was occupied by the final extracts of the cooling batholith.

Rocks of Lode-Zone.

The lode-zone, which traverses the dark brown schist, varies from a few feet to a few hundred feet in width, and consists mainly of a grayish, sugary quartzose rock (greisen) enclosing

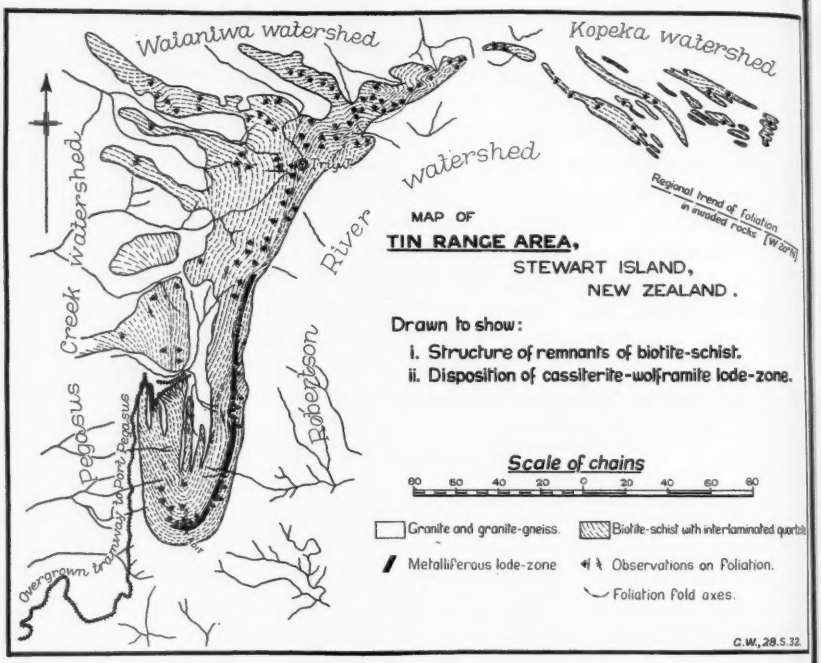

Fig. 2. Tin Range Area, Stewart Island.

an anastomizing series of metalliferous veins only a few of which exceed 6 inches in width. The boundary between the greisen and the schist is sharply defined, but lenticular apophyses appear throughout the neighboring schist, forming eye-shaped distensions in the folia. Only rarely do the metalliferous veins pass from their greisenous host into the surrounding schist. Muscovite is replaced by topaz in the schist near the lode-zone.

Petrological Relationships between Rocks of Lode-Zone.

This brief field description may be expressed in petrological terms. A metalliferous rest-magma was intruded along a narrow zone of weakness in the eastern side of the schist pendant making up the Tin Range. During and after the injection, the rest-magma separated into a gaseous and two liquid fractions:

a. Fugitive pneumatolytic agents that reacted with the biotite-muscovite-schist to form a *biotite-topaz-schist.*

b. A siliceous ichor, containing fluorine and water, that crystallized as a fine-grained *greisen* composed mainly of quartz, topaz, and micas.

c. A liquid metalliferous residue that crystallized as a composite *tin-tungsten lode-system* in the greisen from which it had been differentiated.

In most areas mineralized by the tin-tungsten type of fluids, the mineral lodes are surrounded by an aureole of intensely altered country-rock (greisen) formed by reaction between emanations passing outward from the lode-channels. But the mutual relationships between the rock-types associated with the mineral lodes of Stewart Island are of an *intrusive* rather than a *reactive* nature. The three different rock-types do not represent successive degrees of intensity of alteration of country-rock by emanations from a metalliferous vein; they are successive differentiates (showing mutually intrusive relationships) of a rest-magma intruded along a zone of structural weakness in the roof of the granite.

The mutual relationships between the three rock-types associated with the metallization are shown in Figs. 3 and 4. For the sake of neatness, these drawings are made from samples where the three rock-types occur closely together; it must be borne in mind that the greisen is a few hundred feet wide in most places and is the dominant rock of the lode-zone.

Biotite-topaz-schist.—This rock forms an irregular aureole around the greisen. In thin-section, Fig. 4-C, it is seen to consist of a coarse-grained aggregate of unstrained biotite and topaz.

The mica exhibits pleochroic colors ranging from pinkish brown to straw-yellow. It has been extensively bleached—a process that brought about the relegation of iron-ore to the cleavages and cracks between contiguous crystals. In many cases, the

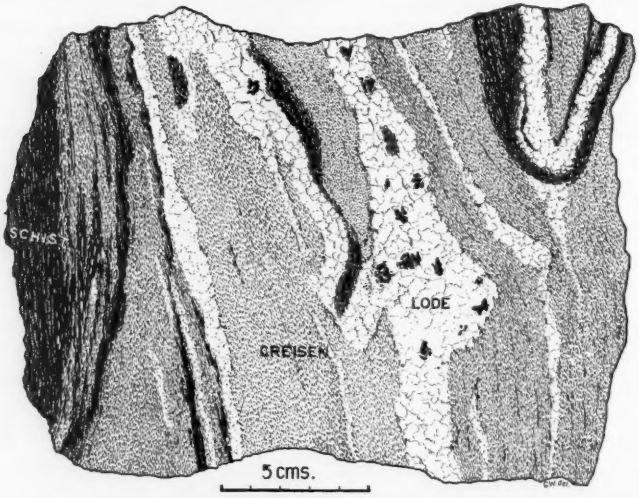

FIG. 3. A typical block from the metalliferous lode-zone.

bleaching has removed all color from the mineral, which can be distinguished from muscovite only by faint brownish haloes around enclosed zircons. The topaz grains are crowded with inclusions, but the margins are mostly clear. The included matter is mainly faint-brown mica, but green spinel (gahnite), quartz, and muscovite were also recognized. The absence of these inclusions from the periphery suggests entanglement during rapid growth, rather than a later alteration such as damouritization.

The entire absence of free silica from this rock suggests pneumatolysis—a presumption that is also supported by the fact that the reaction was sufficiently vigorous to recrystallize the thatched bundles of biotite in the original schist (Fig. 4-D) into large unstrained plates (Fig. 4-C).

Intrusive Greisen.—This is the dominant rock in the lode-zone. It is composed of quartz, topaz, biotite, and muscovite, making up a fine-grained granular aggregate (Fig. 4-B). Quartz is the dominant mineral and rarely incloses delicate hairs of sillimanite. The topaz does not contain inclusions, as does that of the topazized schist. It has presumably crystallized directly from the magma. The small plates of biotite show pale pleochroic colors similar to those of the topazized schist; it is not associated with flecks of iron-ore, but a few zircons were seen. Muscovite is subordinate to biotite; it is probable that small amounts of feldspar, now represented by kaolin, formerly existed in the rock.

The contact between this granular material and the surrounding biotite-topaz-schist shows none of the gradational effects ex-

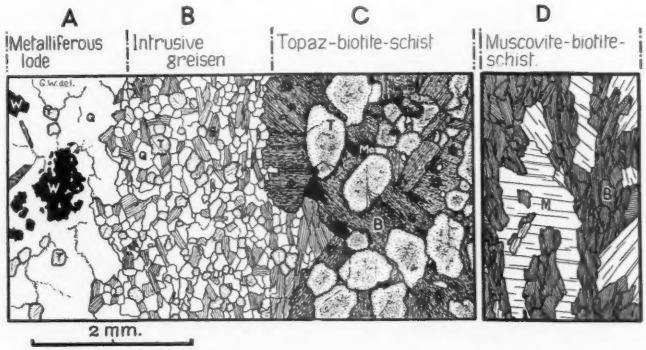

FIG. 4. Drawing of thin-section showing intrusive relationships between the three rock-types **A–C** of the lode-zone. **D** is the unaltered schist. *W*, wolframite; *M*, muscovite; *T*, topaz; *Q*, quartz, *B*, biotite; *Ma*, magnetite.

pectable if the greisen were a replacement type. The fine-grained highly siliceous greisen meets the coarse schist (carrying no free silica) along a ragged contact (Fig. 4). Near the main intrusion of the greisen, lenticular apophyses appear in the schist, forming eye-like distensions in the foliation. Moreover, fragments of the schist appear within the greisen; they are mostly

bent, and the faint gneissic structure of the greisen flows around the bent fragments (right-hand side of Fig. 3). In some places, certainly, a more gradational contact is seen, but sharp contacts are more general.

All these features suggest that the greisen bears an intrusive rather than a reactive relationship to the enclosing schist, and it must be inferred that it was introduced as an aqueous fluorine-bearing siliceous liquid. It does not mark an advanced stage in the metasomatic alteration of the country-rock by pneumatolytic agents passing from a central fissure as do some more typical greisens; such a process would not render the rock fluid so that it could flow past its host-rock. It must be regarded as a petrological entity, formed by advanced magmatic differentiation, and not by post-consolidation pneumatolysis. Just as with most other intrusive bodies, some contamination is in evidence, for in no other way can the large amount of biotite and the presence of sillimanite be explained.

The idea of an intrusive greisen is not novel, nor is there any theoretical objection to such a conception. Greisen is an end-product towards which the whole trend of magmatic differentiation is directed. In some cases equilibrium is attained in the liquid phase and the resulting fluid may be exuded to crystallize as a greisen apophysis of the main mass. But in most cases, the rock freezes, or commences to freeze, before equilibrium is attained, and the fugitive gases maintain the same trend by reaction with the minerals as they crystallize,[1] and, later, with the minerals of the consolidated rock [2] to form typical greisens such as those of Cornwall.

Harker (1895), and also Finlayson (1910), have described a greisen from Grainsgill, that seems to have been extruded as a petrological entity from the Skiddaw Granite, and Scrivenor (1914) has drawn attention to an intrusive "quartz-topaz-rock" at Gunong Bakau. Jones (1916) expresses the belief that the

[1] A process comparable with the "primary pneumatolytic phase" of pegmatite formation of Gevers and Frommurze, 1929, p. 136. (For references, see Bibliography at end of paper.)

[2] The "secondary or principal pneumatolytic phase" of Gevers and Frommurze.

topaz of this rock is derived from feldspar, but Ingham (1930) has recently confirmed the intrusive relationship of the rock. The blending of late magmatic reactions with post-consolidation pneumatolysis is probably responsible for the divergent views taken by these writers.

Metalliferous Lodes.

The metalliferous lodes appear mainly within the greisen, but are occasionally seen in the adjacent topazized schist. They form a composite vein-system of irregular anastomizing veinlets, rarely exceeding 6 inches in width. In places they are contorted (Fig. 3), but the minerals of the contortions show no cataclasis and the twisting is shared neither by the neighboring veins nor by the enclosing greisen; moreover, the veins have a tendency to follow contortions of bent schist inclusions in the greisen. It may be concluded that the lodes were never plane, and that they followed a complex fracture-system, the irregularities of which were due to "the resistance of the rock to plane fissuring" (Read, 1927). Small-scale contemporaneous faulting is in evidence.

Under the microscope, the vein material is seen to be made up predominantly of coarse-grained quartz, with a few delicate hairs of sillimanite, and incipient brownish flakes of biotite; minute prisms of zircon are inclosed in the quartz, and lines of minute indeterminate inclusions pass from one crystal to another.

Topaz occurs in unevenly distributed aggregates in two forms: as groups of large crystals about 2 mm. across with irregular margins, and as aggregates of small geometrically shaped grains isolated in the quartz or filling cracks traversing groups of wolframite crystals. The fine-grained topaz is clear, but the coarse grains are crowded with large numbers of inclusions, among which incipient flakes of biotite, and tiny blebs of wolframite were recognized. It seems probable that the large clouded grains are derived from feldspar or muscovite, whereas the smaller ones have crystallized directly from the vein-solutions. Incipient damouritization is in evidence.

Garnet (spessartitic almandite — 17.30 per cent. MnO) is a

FIG. 5. Paragenesis of metalliferous lodes. *W*, wolframite; *Q*, quartz; *C*, cassiterite; *F*, feldspar; *T*, topaz; *G*, garnet; *M*, muscovite; *B*, biotite.

constant constituent of the mineral veins, and may be even more abundant than topaz. It has a pale plum color and is not zoned. It occurs as large rounded or hypidiomorphic grains, and as irregular forms occupying fractures in the ore-minerals. Rounded inclusions of topaz and quartz are invariably present, and less often, blebs of wolframite.

The masses of wolframite (19.6 per cent. FeO; 4.0 per cent. MnO) that are so clearly visible in hand-specimen are composed of aggregates, the individuals of which rarely show idiomorphic outlines. The margins are rounded in detail and often highly denticulate, and may be surrounded by small euhedra of wolframite, which may represent a redeposition of material corroded by the vein-solutions from the edges of the earlier formed grains. Under strong transmitted light, thin cleavage fragments transmit a deep blood-red light.

Cassiterite is much less abundant than wolframite, and occupies cavities in the surfaces of the wolframite clusters. The pleochroic colors range from grayish brown to brown, and are characteristically blotchy.

Orthoclase was seen in only one slide, where it occupies a crack between two wolframite grains from its close association with muscovite in the same crack, it may be inferred that most of the original feldspar has passed over into muscovite.

Twisted xenocrysts of light and dark mica also appear. Occasional idiomorphic unstrained plates of biotite, and thin hairs of sillimanite point to the incorporation of foreign matter.

Order of Crystallization of Lode-Minerals.—Wolframite and cassiterite were the first minerals to crystallize. The perfect cleavage of the former rendered it susceptible to mechanical disintegration, and each of the gangue minerals occupy parted cleavages and fractures (Fig. 5), the sides of which were to some extent rounded by reaction with the vein-solutions, which later deposited a crop of minute idiomorphs around the periphery of the wolframite clusters; a redeposition that *followed* the crystallization of garnet and topaz. Feldspar formed after the ore-minerals, and was later either muscovitized or topazized, and such as survived was later kaolinized.

Possible Carriers of Tin and Tungsten.

Tourmaline (with or without topaz) is commonly, but not universally, associated with tin deposits. This has given rise to the belief that boron may be a vehicle for the migration of tin. Tourmaline does not occur near the metalliferous lode-zone in Stewart Island. Coombe (Coombe and Groves, 1932) and Simmons (1932) in Uganda, and Wagner (1921) in South Africa, have shown that, although tourmaline is a common mineral in the neighborhood of the deposits they studied, it crystallized before the cassiterite. They therefore conclude that boron was not an active carrier. Wagner (1921) cites Knopf's work in Alaska to show that, under certain circumstances, boron is a carrier. If, however, cassiterite may enter into aqueo-siliceous solution, as hereinafter suggested, there is no reason why the introduction of the tin into the localities could not have been effected through the medium of boron.

Bischoff (1866) long ago suggested that stannic acid is held in solution by alkaline carbonates. But carbonates are characteristic of hydrothermal conditions rather than those of granite contacts, near which most tin deposits are found.

The common paragenetic association of cassiterite and pyrite and arsenopyrite in Malaya, and the infrequency with which topaz is seen, led Jones (1916) to express the belief that tin may be transported as a sulphide or arsenide. The association of cassiterite with these minerals, however, appears to occur in the upper part of the tin zone, or, in other words, where it overlaps the base of the sulphide zone. Gevers and Frommurze (1929) have described a hydrothermal stanniferous vein in which cassiterite occurs with arsenopyrite, pyrrhotite, and pyrite, and suggest that sulphur and arsenic "served as additional aids to solution." These writers do not discuss the reaction by which the metallic sulphide and arsenide yield the metallic oxide.

The scarcity of topaz in close association with tin deposits has also been noted by other observers, who have advanced a variety of hypotheses to explain the origin of the cassiterite. Stheeman (1932), Coombe (Coombe and Groves, 1932), and Simmons

(1933 and postscript to Coombe and Groves, 1932) have described the constant paragenetic association of cassiterite and mica in some stanniferous lodes in Uganda, and Gevers and Frommurze (1929) have noted similar characteristics in stanniferous pegmatites in South-West Africa. The association led Coombe and Simmons to the conclusion that the tin was introduced as an alkali stannate, which reacted with alumina (presumably from the wall rocks) and silica to produce cassiterite and muscovite. Simmons suggests the following reaction:

$$K_2O.SnO_2.3H_2O + 6SiO_2 + 3Al_2O_3 = K_2O.2H_2O.3Al_2O_3.6SiO_2 + SnO_2 + H_2O.$$

But there seems no necessity to postulate such a medium for the migration of tin, for, as pointed out by these writers, the micas were not analyzed and may contain some fluorine. This point has also been raised by Groves (1933), who submits reasons for believing the mica to be fluoriferous. Finlayson (1910) determined 0.92 per cent. of fluorine in the muscovite of the Grainsgill greisen, and there is 2.23 per cent. in the biotite adjacent to the lodes in Stewart Island.

Rôle of Fluorine in Migration of Tin and Tungsten.

It is generally believed, however, that fluorine is the principal carrier of tin and tungsten.

Cassiterite.—Daubrée's (1879) reaction figures prominently in the literature of tin deposits:

$$SnF_4 + 2H_2O = SnO_2 + 4HF.$$

J. H. L. Vogt (1926) has pointed out that this equation is reversible, the reaction from left to right being promoted by an increase in the H_2O/HF ratio, a condition that may be brought about by the escape of HF into the surrounding rocks or by the liberation of water resulting from the dissociation of such a compound as $H_2SiO_3.nH_2O$ to form quartz.[3] The extraction of SnO_2 is promoted by the reciprocal conditions, thus account-

[3] A *gas* phase may develop by this dissociation only under moderate and low pressure conditions (Jakob, 1919).

ing, according to Vogt, for the presence of cassiterite only in the latest emanations from granitic magmas. Wagner (1921) suggests that stannic fluoride may react with silica to form SiF_4 and SnO_2.

Wolframite.—The frequent paragenetic association of wolframite and cassiterite suggests that tungsten, also, is carried as a fluoride (*e.g.,* Rastall, 1918). From what source is the (FeMn) radicle derived? Magmatic differentiation effects a progressive decrease in the MgO/FeO ratio, and, later, a decrease in the FeO/MnO ratio in successive liquid residues. The RO content of the final residuum may, therefore, be expected to consist mainly of FeO and MnO, with a tendency for the latter to increase at the expense of the former. This trend has also been observed by Vogt (1906).

But there is also the possibility that the FeO may be obtained from the country-rocks by leaching; the baueritization of biotite under pneumatolytic conditions is well known (Brammall and Harwood, 1923). The bleaching of the biotite adjacent to the metalliferous lode-zone in Stewart Island may be indicated in the following form:

$$\underset{\text{(biotite)}}{\text{lepidomelane} + \text{phlogopite}} = \text{muscovite} + \text{phlogopite} + 4FeO + 3Fe_2O_3.$$

The original biotite thus splits up into a magnesia-rich biotite, muscovite, and iron-ore. The released iron may remain in the system (as in Fig. 4-*C*, where it occupies cleavages and intercrystal spaces) but more often is lost, and may possibly contribute to the RO of the wolframite of contiguous veins.

Physical State of Fluorides at Granite Contacts.

The general belief that tin and perhaps tungsten are introduced as fluorides is based on Daubrée's (1879) experiment and upon the paragenetic association of cassiterite and wolframite with fluorine-bearing minerals (notably topaz and mica). This circumstantial evidence may be checked by inquiring if the mode of occurrence of cassiterite and wolframite is in keeping with what we may suppose, by using such physical data as are available, to be

the physical condition of the fluorides of tin and tungsten under the conditions of a granite contact; the mineral products of gaseous and liquid reaction should present distinguishable field characteristics.

Wright and Larsen (1909) have shown from a study of quartz, that pegmatites crystallize between 700° and 550°; these temperatures have been confirmed by the recent experiments of Goranson (1932). As Jones (1920), Derry (1931) and others have demonstrated that pegmatites may grade upwards into tin-tungsten lodes, it may be inferred that the temperature of formation of such lodes is, in general, less than 550°.

The pressures at granite contacts must vary considerably from one mass to another; some intrusions are deep-seated, whereas others, such as the Stewart Island batholith, may reach comparatively shallow depths. But, as the late emanations from granitic magmas mostly occupy tension joints and gashes in peripheral fracture-systems, there is no justification for believing them to have existed under very high pressure.

At atmospheric pressure, the boiling point of SnF_4 (according to Vogt, 1926) is 705°, and that of WF_6 is 19.5°. The pressure of a granite contact would raise these boiling points, but they would be reduced in the presence of other gases. What substances are likely to be gaseous under such conditions? Vogt (1926) believes the critical temperature of water does not rise above 425°. The boiling point of SiF_4 is —65°. Although van Nieuwenburg and Blumendal (1931) reported the volatilization of silica in supercritical steam,[4] Morey (1932) repeated the experiments and observed no substantial migration of silica. However, some at least of these substances must have been volatile under the conditions of a granite contact. Such of them as were volatile, would have a depressing effect on the boiling point of SnF_4 that would probably be of a higher order than the increase of boiling point produced by the pressure-increase over the normal pressure (for which the boiling points are known). The boiling point of SnF_4 would thus be reduced from 705° to

[4] SnO_2 is not volatile under these conditions.

a lower figure, but it is by no means certain that SnF_4 would be in a gaseous state, a doubt that has previously been expressed by Lindgren (1928, p. 724) in discussing the origin of cassiterite: "In the absence of exact data as to the behavior of the solutions at this critical temperature, the insistence on pneumatolysis has, however, little value."

Wolframite and Cassiterite not always Deposited by Gaseous Interaction.

By a not infallible physico-chemical argument, the result is obtained that SnF_4 may not be gaseous at a granite contact. Can this doubtful result be substantiated by field-evidence?

If cassiterite and wolframite were precipitated directly from the gas-phase, their field-occurrence should present features consistent with those of gaseous reaction: they should invariably appear either as replacement products, or as euhedra lining vugs; and in the latter case they should be found abundantly in the adjacent country-rock, for gases may be expected to migrate readily. The mere growth of euhedra in vugs is not, *per se*, proof of pneumatolytic deposition.

They do actually occur under both these circumstances. Euhedral growths of wolframite and cassiterite in vugs are frequently found, and they also occur as replacement minerals in metalliferous greisens such as those of New South Wales (Cotton, 1909), Cornwall (MacAlister, 1908), and South Africa (Wagner, 1921). In these cases pneumatolytic deposition is possible. But cassiterite and wolframite commonly occur in siliceous veins and as primary minerals in pegmatites, unattached to the walls, and absent from the country-rock. These features are exhibited by the veins of Stewart Island and by many of the lodes described by Jones (1920 and references therein) from Malaya. In fact, stanniferous veins are as common or even commoner than stanniferous greisens. The ores in many of these veins are clearly not a replacement product, and as they usually show no tendency to grow from the vein-walls in euhedra and are commonly absent from the country-rock, it may be surmised that the metals did

not come into the position they now occupy in the veins in gaseous combination.

There are several reasons for believing the Stewart Island veins were not deposited from the gas phase:

1. The metalliferous minerals are restricted to the actual veinlets. If they had resulted from interactions between gases, one would expect the gaseous metalliferous fluorides to have permeated the country-rock.

2. Vein-banding and comb-structure do not appear. There is no evidence pointing to the origin of the veins by successive depositions on the sides of fissures; the crystal structure of the veins indicates deposition from solutions that were pressed into and filled the fracture-system.

3. Feldspar crystallized in the veins (but was later almost entirely muscovitized); it is inconceivable that a feldspar should result from gaseous interaction.

4. The veins contain no vugs.

5. Garnets occur in the lodes; they have a composition comparable with those found in pegmatite dikes in the district.

6. Several of the constituent minerals show marginal resorption; this is particularly noticeable with garnet and wolframite, and in the case of the latter, some of the resorbed material was redeposited in a quartz medium as minute euhedra about corroded cleavage faces.

Migration of Ores in Solution.

The conception of transfer of metals in chemical combination with light elements is deep-rooted in metallogenetic studies, but plays only a minor rôle in the literature of petrogenesis. The minerals that are the special concern of igneous petrologists migrate from place to place in siliceous solution (in magmas), and when they reach a temperature zone that demands their crystallization, they are deposited one after another. Economic geologists have to contend with an additional difficulty, for this simple law of magmatic differentiation is complicated by gaseous transfer of metallic radicles. But it is a difficulty that has been overestimated.

Although there must remain no doubt that the concentration of valuable metals in the final ichors of magmas may be assisted by combination with the lighter elements, the conception of volatile "carriers" would seem to be given disproportionate weight in discussions on the migration of valuable metals from batholithic margins. Recent physico-chemical work is revealing an unsuspectedly high *solubility* of ores in hot aqueous and aqueo-siliceous solutions.

This theme may be adapted to the special problem of this paper —the migration of tin and tungsten. Although there seems no good reason to doubt that tin and tungsten are brought to granite margins in combination with fluorine, there is field-evidence from many mining fields pointing to their having also been transferred in some other manner after the metallic radicles had been released from their carriers. Aqueo-siliceous solution offers an attractive alternative to chemical combination.

Some evidence may be summarized suggesting such a mode of transfer; although several of the points mentioned below are individually inconclusive, they form, together, a strong argument for deposition from solution:

1. Fluorine is rare in the vicinity of many of the "quartz-vein deposits" of Jones (1920) and the Tavoy veins of Campbell (1919).

2. The ores in some of the stanniferous deposits of Uganda and South Africa are in siliceous veins invading previously tourmalinized country-rock; if the tin had been in combination with boron right up to the time of its deposition as cassiterite, the cassiterite should be found in intimate paragenetic association with tourmaline.

3. Peripheral resorption and redeposition of wolframite imbedded in quartz were observed in Stewart Island. This suggests that siliceous solutions are capable of dissolving wolframite.

4. Many tin-tungsten lodes exhibit no regular banding, comb-structure, or vugs; the ore and gangue minerals have clearly consolidated from a solution that was forced into a fracture and occupied the whole of the space.

5. The existence of regular banding, comb-structure, and vugs is not evidence for deposition on the side of a fissure from the gas-phase. Gases so readily diffuse that it is unbelievable that they could press apart the walls of a fissure. Such vein features are more readily explained by rhythmic deposition from a solution that was forced into the fracture—the walls being held apart by the pressure of the solution. The vugs represent the aqueous residue from the crystallization of the vein-minerals. This residue maintained the pressure until later earth-movements brought the vein-system into a shallower zone and permitted its escape.

6. The common occurrence of euhedra of ore-minerals entirely surrounded by siliceous gangue is a clear indication of deposition from solution.

7. Wolframite and cassiterite have been reported as primary minerals in pegmatites—pegmatites commonly grade upwards into metalliferous veins.

Order of Deposition of Cassiterite and Wolframite.

The order of deposition of cassiterite and wolframite in Stewart Island was not ascertained with certainty, but wolframite was probably the first to form. The extensively investigated tin-tungsten province of Malaya gives information on this matter. After an extensive field-study of the tin-tungsten lodes of Malaya, Jones (1920) maintained that where tin and tungsten occur together in the same mineralizing solution, the deposition of tin invariably precedes that of tungsten. From this observation, he deduced that "... where the cassiterite and wolframite occur together in the same field, the ratio of tin to tungsten is less, *in general,* in the lodes farthest from the granite than in those traversing the granite." In the discussion that followed the presentation of his paper, it was agreed that, in general, wolframite travelled farther from the granite than cassiterite, but several exceptions were mentioned. Some of these could be explained by vein-fracturing and redeposition within a rising geotherm, but in other cases that were cited, the cassiterite had

actually formed on the surface of wolframite from the same solutions.

The following explanation of these anomalies does not invalidate the general trend of Jones's remarks. The main reason for the depth-interval between cassiterite and wolframite is no doubt due to the wide discrepancy between the boiling points of SnF_4 and WF_6 (705° and 19.5° at atmospheric pressure).

Cassiterite is probably produced in accordance with Daubrée's reaction; and it is possible that wolframite is produced by the action of H_2O on WF_6 in the presence of iron and manganese. Reactions in such reactive systems are governed by the law of mass action. If the ratio W/Sn were large, wolframite would be deposited before cassiterite; if it were small, these minerals would be precipitated in the reverse order. The physical state of the reacting substances has no influence on the actual reactions, but it may influence the proportion of the reacting substances present in so far as gases are more readily transferable than liquids, and the low boiling point of WF_6 would tend to eliminate W from the system after the liquefaction of SnF_4.

It has been shown that the mode of occurrence of cassiterite and wolframite in some veins suggests that they may also crystallize from siliceous solutions. Now, deposition from solution is governed not only by the relative proportions of the minerals present but also by their relative melting points. Derry (1931) and Ruff (cited by Derry) have shown experimentally that the melting point of SnO_2 is over 1600°, and wolframite undoubtedly melts at a lower temperature. Consequently, cassiterite would, in general, be the first of these minerals to crystallize from solution, but if wolframite were present in preponderating amount, its precipitation would precede that of cassiterite in accordance with eutectic principles. It may be expected, then, that wolframite would be the dominant mineral in those cases where its deposition has been succeeded by that of cassiterite; this seems to be the case in Stewart Island.

To sum up, cassiterite is in general deposited at lower levels than wolframite. Liquids are less mobile than gases, and the

higher boiling point of SnF_4 retards its retreat from the granite margin. But a reversal of this normal order of deposition may be brought about by the dominance of W in a reactive system containing SnF_4, WF_6, and H_2O, or by the preponderance of wolframite in a eutectic system containing SiO_2, SnO_2, and $(FeMn)WO_4$.

Summary of Conclusions.

1. A granite batholith penetrated into the zone of directed pressure, and invaded a biotite-muscovite-schist. A late magmatic residue was intruded along a zone of weakness in the schist. During and after injection, this residue gave rise to (*a*) fugitive gases which permeated the adjacent biotite-muscovite-schist and changed it to a biotite-topaz-schist; (*b*) a quartz-mica-topaz rock that may be called a greisen; and (*c*) a metalliferous residue that occupied a fracture-system in the greisen to form a wolframite-cassiterite lode-system.

2. The quartz-mica-topaz rock (greisen) contains no residual structures, and there seems no doubt that its minerals were precipitated directly from the residual magma. It is regarded as one of those rare cases where magmatic differentiation achieves its logical end-product in the fluid phase; most greisens are formed by the pneumatolysis of the consolidated mother-rock, but the advanced product of such reactions is not always petrographically distinguishable from the late differentiation-product.

3. It is believed that the extraction of tin and tungsten from granitic magmas is effected by HF, that they are transported to granite margins as fluorides (possibly in a gaseous state), and that the oxide and tungstate are formed at the granite contacts by dissociation of the fluoride in the presence of H_2O. In some cases, cassiterite and wolframite may remain where deposited, but in general they are further transported and finally crystallize in accordance with the principles of eutectic crystallization. This transport may remove them from the sphere of fluorine-bearing minerals, and thus afford an explanation for the scarcity of fluorine-bearing minerals in the neighborhood of some tin-tungsten deposits.

4. In those cases where cassiterite and wolframite are deposited by chemical reaction, the law of mass action would demand the precipitation of wolframite before cassiterite if it were present in greater amount. But in cases of deposition from solution, wolframite would crystallize first only if present in preponderating amount.

Acknowledgments.

For encouragement and advice during the preparation of this paper, the writer is indebted to Professor Benson and Professor Park, of Otago University, and Professors Boswell, Brammall, and Cullis, Dr. Jones, and Dr. Harwood, of the Imperial College of Science and Technology. He gratefully acknowledges financial assistance given by a Government Grant administered by the New Zealand Institute, the Lubecki Scholarship of Otago University, and the Post-Graduate Scholarship in Science of the University of New Zealand.

The paper forms a section of a thesis submitted for the Degree of Doctor of Philosophy of London University. A discussion of other phases of the geology of Stewart Island will appear in forthcoming papers.

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RECONNAISSANCE OF BURIED RIVER GORGES BY THE EARTH RESISTIVITY METHOD.[1]

STANLEY W. WILCOX AND G. M. SCHWARTZ.

INTRODUCTION.

IN connection with studies of the topography of the rock surface beneath the glacial drift in the Minneapolis-St. Paul area in Minnesota, it was found that available well records left gaps in the data, particularly as to the location of the sides of the buried river gorges. As a rule, geophysical methods are too expensive to use for general work of the sort undertaken by State geological surveys, but it was felt by the junior author that some rapid method should be available. The problem was therefore suggested to Mr. Wilcox, at the time a graduate student in Physics at the University of Minnesota, as a subject for a Master's thesis. Emphasis was placed on the development of a procedure which would be relatively rapid and inexpensive, corresponding somewhat to the usual expense of geologic mapping of an area.

The writers are indebted to Dr. O. H. Gish, inventor of the Gish-Rooney Earth Resistivity apparatus, for permission to use his circuit in these experiments; also to the Minnesota Geological Survey for field expenses, and to the Department of Physics of the University of Minnesota for the loan of part of the equipment used in construction of the instrument.

GEOLOGIC SETTING.

The region about Minneapolis and St. Paul has been heavily glaciated, and thick deposits of glacial drift cover the rock surface except locally along the Mississippi, Minnesota, and St. Croix Rivers. The glacial geology is discussed at length by Sardeson [2]

[1] Published with the permission of the Director of the Minnesota Geological Survey.

[2] Sardeson, F. W.: U. S. Geol. Survey Geol. Atlas, Folio 201, 1916.

and by Leverett,[3] and need not be summarized here. The pre-glacial topography of the region was probably a mature type somewhat like that found today in southeastern Minnesota along the Mississippi river. Consequently, beneath both cities and in adjacent areas there are several deep pre-glacial gorges filled with glacial deposits. Those found in Minneapolis have been described in detail by Soper.[4] Several of these gorges have a surface expression resembling the one that trends through Minneapolis and is occupied by Lake of the Isles, Lake Calhoun, and Lake Harriet. Other gorges have no surface expression but are heaped high with terminal moraine deposits, as is conspicuously shown by the gorge that trends west through Wescott, south of St. Paul (see Folio 201). Several wells show this gorge to be buried beneath 400 feet of glacial material. Within the cities, these gorges are important in foundation work and in the construction of trunk sewer tunnels. Over a large area in and surrounding the cities, they are of great importance in connection with ground-water problems, particularly in regard to questions of supply of artesian water which is widely used throughout the region. For this reason more information on the subsurface rock topography has been in demand and the Minnesota Geological Survey is preparing a report covering these special features as well as the general geology of the Metropolitan area. Some geologic problems for which geophysical data were utilized are described below.

SELECTION OF METHOD.

All geophysical methods depend for their operation on physical discontinuities in the crust of the earth. These discontinuities, whether they be of density, elasticity, electrical resistance, or of magnetic susceptibility, will correspond in a general way with variations in the structure and the petrology of the beds. For example, if the glacial cover or weathered mantle has an average

[3] Leverett, Frank: Quaternary Geology of Minnesota and Parts of Adjacent States. U. S. Geol. Survey Prof. Paper 161, 1932.

[4] Soper, E. K.: Buried Rock Surfaces and Pre-glacial River Valleys of Minneapolis and Vicinity. Jour. Geol., vol. 23, pp. 444–460, 1915.

specific electrical resistance differing from that of bedrock, an electrical survey would be indicated as a means of determining the depth to bedrock.

The local geology and a study of the previous work done on similar areas afford the criteria for deciding what geophysical method could be used for any proposed subsurface exploration. For the present problem, the cost of the survey rules out both the torsion balance and the seismograph; these methods, although they might yield the desired information, are not economical in their present state of development for shallow investigations.

A magnetometer does not seem applicable for measuring the thickness of the drift except possibly in northern Minnesota, where the drift is in contact with pre-Cambrian rocks. It is probable that the magnetic differences between the drift and the consolidated sedimentary beds of the Minneapolis-St. Paul area are small and erratic, and correlation of these differences with structure might be uncertain or impossible.

The geo-electrical method appears to be the logical one for shallow investigations. The first cost is reasonable because most of the equipment can be assembled from standard laboratory instruments. Furthermore, the field technique is simple, rapid, and inexpensive.

Electrical prospecting or surveying is based on the variations of some electrical property of the crust of the earth. A buried conductor will distort an alternating current field induced in its vicinity. Earth-resistivity measurements on the surface are functions of the conductivity and the position of the geologic beds included in the effective volume through which the current flows from one earthed electrode to the other. High frequency geo-electric methods are effective for locating buried conductors, and thus are of chief application to ore-deposit problems. The earth-resistivity method, on the other hand, is particularly suitable for structure surveying where the beds are not complexly folded and where the electrical discontinuities can be correlated with key geological horizons. This method seems best for locating the edges of the buried gorges because at these points appreciable

electrical resistance anomalies would be expected. Furthermore, the moist zone above the electrically resistant Platteville limestone would appear to be a characteristic electrical horizon. With these considerations in mind, the Gish-Rooney Four-Electrode Method was selected as the most suitable one for the investigation described in this paper.

REVIEW OF HISTORY OF METHOD.

The first earth-resistivity experiments were made by Professor Conrad Schlumberger, of the School of Mines in Paris, at Fierville-la-Compagne, Normandy, in 1912. He published this work in 1920,[5] and it is still an authoritative treatise of earth-resistivity measurements in spite of the work that has been done since that date.[6]

Little attention was paid to earth-resistivity work in this country until 1925, when Dr. O. H. Gish, of the Department of Magnetism, Carnegie Institution, developed what is now known as the Gish-Rooney Earth-Resistivity Method, based on the theoretical work of Frank Wenner,[7] and obtained a patent on the instrument (U. S. patent No. 1,813,845). Mr. Rooney, of the same institution, tested the instrument near Washington, D. C.; at Waterloo, Western Australia; and at Ebro, Spain. The first practical application of this geophysical method was made in the Michigan copper country in 1927, where considerable research was carried out to determine the apparent resistance of the country rock associated with native copper ores.[8]

Irving B. Crosby [9] and E. G. Leonardon [10] appear to have been

[5] Schlumberger, C.: Etude sur la Prospection Electrique du Sous-Sol. Gauthiers Villars, Paris, 1920.

[6] Leonardon, E. G., and Kelly, Sherwin, F.: Some Applications of the Potential Methods to Structure Studies. A. I. M. E., Geophys. Prosp., 1929, pp. 180–199.

[7] Wenner, Frank: A Method of Measuring Earth-Resistivity. U. S. Bureau of Standards, Science Paper 258, 1915.

[8] Hotchkiss, W. O., Rooney, W. J., and Fish, James: Earth-Resistivity Measurements in the Lake Superior Copper Country. A. I. M. E. Geophysical Prospecting, 1929, pp. 51–65.

[9] Crosby, Irving B., and Kelley, Sherwin, F.: Electrical Subsoil Exploration and the Civil Engineer. Engineering News-Record, February 14, 1929, pp. 270–273.

[10] Leonardon, E. G., and Kelly, Sherwin, F.: Some Applications of Potential Methods to Structure Studies. A. I. M. E. Geophysical Prospecting 1929, pp. 180–199.

the first to apply the earth-resistivity method to problems of civil engineering. They investigated a dam site on the upper Connecticut River in 1928. Leonardon [11] states that more than twenty investigations at dam and tunnel sites have been made by his company in the United States and Canada. He also outlines the results of his electrical work at Bridge River Station, British Columbia, where an earth-resistivity survey was used to predict the geological conditions that would be encountered in boring a tunnel. He says: "The discrepancies between the predictions of the electrical survey and the results of the underground exploration are small," and further, "the correspondence between the electrical resistivities and the mechanical properties of rocks is remarkable." In the same article he describes some exploration work on the St. Lawrence River, near Morrisburg, Ontario, Canada, where the drift-limestone contact was contoured with an average error of less than 3 per cent.

Cragmile, Kelly and Low [12] describe some interesting small-scale experiments.

M. King Hubbert [13] found that the Megger was particularly useful for prospecting for gravel, and K. Kurtenacker,[14] of the Wisconsin State Highway Department, reports he is able to differentiate between unconsolidated beds and thus to classify soils for highway construction projects. F. V. Reagel [15] of the Missouri State Highway Department also reports that the method is used as a regular part of cut classification surveys.

THEORY OF EARTH-RESISTIVITY MEASUREMENTS.

If four earthed electrodes are collinear and equally spaced (Fig. 1), an electric current of intensity I flowing between the

[11] Leonardon, E. G.: Electrical Exploration Applied to Geological Problems in Civil Engineering, A. I. M. E. Tech. 407, 1931.

[12] Cragmile, Wm. B., Kelly, Sherwin, F., and Low, Bula: Applying Megger Ground Tester in Electrical Exploration. Reprint, A. I. M. E. New York, 12 pp., 1931. ("Megger" is a trade name for a portable adaptation of the Gish-Rooney Circuit.)

[13] Hubbert, M. King: Results of Earth-resistivity Survey on Various Geologic Structures in Illinois. A. I. M. E. Tech. 463, 1932 (23 pp.).

[14] Personal correspondence with S. W. Wilcox.

[15] Idem.

outside electrodes P_1 and P_4 will produce a difference of potential v between the inner electrodes p_2 and p_3. Frank Wenner [16]

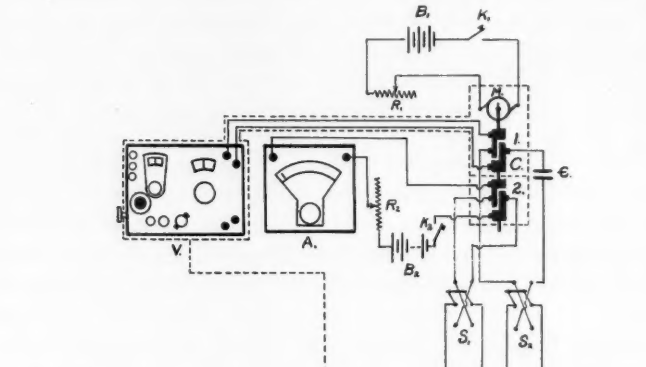

FIG. 1. Generalized diagram of the Gish-Rooney earth-resistivity instrument.

showed that the specific resistance of a homogeneous and isotropic medium of infinite extent is $\rho_a = 2\pi aV/I$ ohms per cubic centimeter if the electrode separation "a" is measured in centimeters.

Only in an ideal case can the surface of the earth be considered a uniform conductor of electricity, and thus the resistance (ρ_a) as determined by Wenner's formula is not a specific constant, but

[16] Wenner, Frank: A Method of Measuring Earth-Resistivity. U. S. Bureau of Standards, Science Paper 258, 1915.

an apparent value depending on the variation of the electrical conductivity with depth and lateral extent. The specific resistance ρ_a (resistance per unit volume) of any small uniform earth section may vary over an extremely wide range of values, from a very small fraction of an ohm per cubic centimeter for native metals and some sulphide ores, to millions of ohms per cubic centimeter for pure quartz and rock salts. Such sedimentary rocks as sandstones and limestones may differ by several hundred per cent. or more even when in contact with each other in the geologic section. The actual flow of electricity is probably electrolytic, the freedom of movement of the ions being largely determined by the water content of the rocks or soils, the salts in solution, the continuity of the rock pore-space, and the mineralized constituents.[17]

Fortunately, the earth-resistivity method of surveying does not depend on absolute resistance values, which may vary greatly even for beds of otherwise uniform physical properties, but only upon the average differences encountered in passing from one bed to another. The total effective earth section through which the current passes is roughly hemisperical, and for the ideal case of an isotropic and homogeneous medium, three-tenths of the total current penetrates below a point equal to the electrode separation. As an analogy the lines of flow of the current through any vertical section including the electrodes are identical with the lines of magnetic flow between the poles of an ideal magnet. When a discontinuity exists, as shown on the instrument diagram, the presence of the lower bed will distort the lines of current flow causing an abnormal distribution of the equipotential lines on the surface.

Gish and Rooney discovered the empirical rule that the depth "h" to the second layer of a two-layer geologic problem is a simple function of the apparent resistance and the electrode spacing. As the electrode spacing "a" is increased, keeping the center of the electrode configuration fixed (for a depth profile), the volume distribution of the current will include more and more

[17] Sundberg, Karl: Effects of Impregnatic Waters on Electrical Conductivity of Soils and Rocks. A. I. M. E., Geophysical Prospecting, 1932, pp. 367–391.

of the second layer, and when the value of "a" equals the depth h, the variation of ρ_a will be a maximum. These inflection points are marked E. H. (electrical horizons) on the curves and can be used to determine the depth of the second layer. This simple rule for finding depths is not strictly valid; it must be used with caution, and carefully checked under control to see how it correlates with well records. It seems, however, to be surprisingly accurate where the electrical discontinuities are small, and sufficiently accurate where they are not small if $h = a - C$, where C is a constant for the particular investigation to be determined by test profiles.

Tagg [18] and others have shown that no theoretical basis can be found for Gish-Rooney's method of depth determination, stating that smooth curves should be expected. Notably Tagg, Ehrenburg, and Watson [19] have done considerable theoretical work calculating what the relation of ρ_a and "a" should be for idealized sections of two or more horizontal layers. Undoubtedly their exact mathematical method should be used where the geological structure can be so simplified, and where the empirical method breaks down. The writers feel, however, that the remarkable correlation obtained by the empirical method justifies its use, especially since the assumptions necessary for the application of a more refined mathematical method often fail to give good results.[20]

For step-traverses where the electrodes are progressively moved forward, the depth of the lower layer is proportional to the apparent resistance, providing depth is less than the electrode separation. (See Tagg, above.)

[18] Tagg, G. T.: The Earth-Resistivity Method of Geophysical Prospecting: Some Theoretical Considerations. Mining Magazine, London, vol. 43, no. 3, 1930, pp. 150–158.

[19] Ehrenburg, D. O., and Watson, R. S.: Mathematical Theory of Electrical Flow in Stratified Media with Homogeneous and Isotropic Layers. A. I. M. E. Tech. 400, 1931 (18 pp.).

[20] Hubbert, M. King: Earth-Resistivity Survey in Illinois. Eng. and Min. Jour., No. 3, 1933, pp. 142–143.

The Instrument.

A radio "B" battery, B_2 (Fig. 1), tapped to give voltages from 2 to 135 volts, is the source of the earth current and is regulated by the rheostat R_2. This current of intensity I is measured by a multi-range milliammeter A with scale divisions, 0–15, 0–150, and 0–1500 milliamperes, and then passed into the commutator C-2 where it is alternated and sent into the earth at p_1 and p_4. The potential difference v between p_2 and p_3 is determined by a Leeds and Northrop field type potentiometer (V in diagram). The scale range on this instrument is 0–16 and 16–64 millivolts. A rectifier C-1, similar to C-2, keeps the current in the same direction as its flow through the milliammeter, thus making possible the use of direct current meters. The condenser ϵ is of 20 micro-farad capacity. Its use tends to steady the galvanometer needle on the potentiometer.

The potentiometer, potential leads, and the rectifier are shielded from the current circuit, the whole being earthed at G, the neutral point.

The commutator is driven by a six volt auto-horn motor, the speed of which is controlled by a two ohm rheostat R_1. This drive was found entirely satisfactory, the motor maintaining constant speeds through a range of 1000 to 2500 r.p.m.

The commutator is of special design. Mr. C. H. Dane of the Physical Laboratory of the University of Minnesota constructed it from plans furnished the authors by Dr. Gish. The alternator and rectifier are so timed that the transient currents caused by the make and break of the current alternator are eliminated from the potential circuit. The interval of time the current circuit is broken must be taken into consideration in calibrating the milliammeter. This correction, called the commutator factor, varies slightly with the speed of the commutator. It is thus important to keep the motor running at constant speed during any given set of observations. For this purpose a speed indicator would be useful; however, the ear soon becomes accustomed to small variations in speed.

The Gish-Rooney Earth-Resistivity Method is essentially a

direct-current one. The frequency of the current sent to the electrodes $p_1 - p_4$ (from 10 to 30 cycles per second), is low enough to exclude an appreciable inductive and capacitative effect, providing the current and potential cables are kept separate. The flow of the current must be alternated to avoid polarization effects at the electrodes and to eliminate stray natural or artificial earth currents.

Small-Scale Experiments.

To test and demonstrate the instrument, some laboratory experiments were conducted in a tank filled with water. These experiments are of value because they predict what would happen under similar relative geo-electrical conditions when the scale is enlarged to practical field dimensions. The experimental tank, eighteen inches by twenty-four inches, made of limestone, was filled with 8 inches of tap water, and in it were placed conductors and insulators in various positions.

Curve *A*, Experiment 1, Fig. 2, shows the effect on the ap-

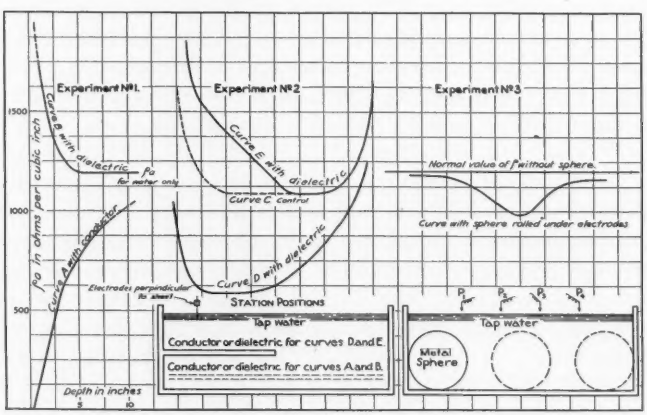

FIG. 2. Curves showing resistivity data obtained from small-scale laboratory experiments.

parent resistance ρ_a when a conductor is placed at various depths in the tank. Curve *B* on the same plate shows a similar experi-

ment for a slab of marble. The position of the electrodes and their separation ("a" = 4 inches for all the small-scale experiments) were kept constant. It will be noticed that ρ_a was 1200 ohms per cubic inch with only water in the tank. When a slab of marble was placed in the bottom of the tank, no change in ρ_a took place, as would be expected from the fact that the tank is made of the same material. When the slab was raised (Curve B), the first noticeable change in ρ_a occurs for a depth of 5¼ inches. For small depths, or better, for small ratios of h/a the apparent resistance increases very rapidly with decrease in the depth of the slab. For curve A with a sheet of iron substituted for marble slab the apparent resistance decreases very rapidly with decrease in depth of the conductor. It is obvious that, if curve A were extrapolated, and a very large tank were used, this curve would approach curve B asymptotically.

Curve C, Experiment No. 2, is the control curve for this experiment. This graph shows the effect of the side walls of the tank when the electrodes, relatively fixed, were moved across the tank from one side to the other. Curve D was obtained in the same manner as Curve C except that a sheet of metal was placed half way down, in and to one side of the tank. Curve E is a repetition of the previous experiment with a slab of marble substituted for the conductor. Experiment No. 2 indicated what should be expected when a step-traverse crosses a horizontal discontinuity.

Buried conductors are often of considerable interest either from the view of their disturbing influence on structure work or for their possible economic value. Experiment No. 3 shows the nature of the apparent resistivity curve when the electrodes were kept fixed and the sphere rolled under them.

The results of these experiments can only be interpreted qualitatively. For quantitative results a large tank should be used. An interesting account of some experiments conducted with geological structure models is given by Joel H. Swartz in his article "Resistivity Measurements upon Artificial Beds." [21]

[21] Swartz, Joel H.: Resistivity Measurements upon Artificial Beds. U. S. Bur. Mines, Inf. Circ. 6445, Feb., 1931 (9 pp., 14 figs.).

Field Data.

Depth Profiles.—The first field test of the instrument was made on the University of Minnesota Campus. The site selected was the Old Parade grounds, where the depth of the Platteville limestone is moderate and accurately known, and where no conducting pipes are present to interfere with the test.

D. P. No. 1 and D. P. No. 2 in Fig. 3 are the curves obtained from the initial tests. These curves are known as depth profiles (D. P.), a vertical earth-resistivity investigation in which the center of the electrode configuration remains fixed while the electrode separation ("a") is increased for each reading of ρ_a, so as to include a larger effective electrical volume and thus greater depth. For interpretation, ρ_a is plotted against "a."

Depth profile No. 1 indicates that the surface material was fairly conducting, but as "a" is increased ρ_a increases to 1000 ohms per cubic foot at "a" = 20 feet when glacial gravels were included in the electrical section. Beyond "a" = 20 feet the decrease is rapid, as the water table or conducting clays are now included in the section. The more resistant Platteville limestone flattens out the curve, and beyond "a" = 50 feet the curve is a straight line. The electrical depth of the bedrock appears to be 50 feet from these profiles.[22] The actual depth from drilling records is 52 feet.

D. P. No. 2 was run near No. 1 to check it by another method. For No. 1 Wenner's electrode arrangement was used, as for all subsequent work. For No. 2, however, a different electrode arrangement was used. In this case the two outer electrodes P_1 and P_4 remained fixed and at a considerable distance apart (300 feet), while the inner electrodes were moved away from one of the current electrodes.[23]

The field technique for this electrode arrangement is more

[22] Crosby, Irving B., and Kelly, Sherwin F.: Electrical Prospecting Applied to Foundation Problems. A. I. M. E., Geophysical Prospecting, 1929, pp. 199–211. See especially pp. 204–208.

[23] Heiland, C. A.: Advances in Technique and Application of Resistivity and Potential Drop-ratio Methods in Oil Prospecting. Bull. Amer. Assoc. Petr. Geol. vol. 16, No. 12, p. 1277, 1932.

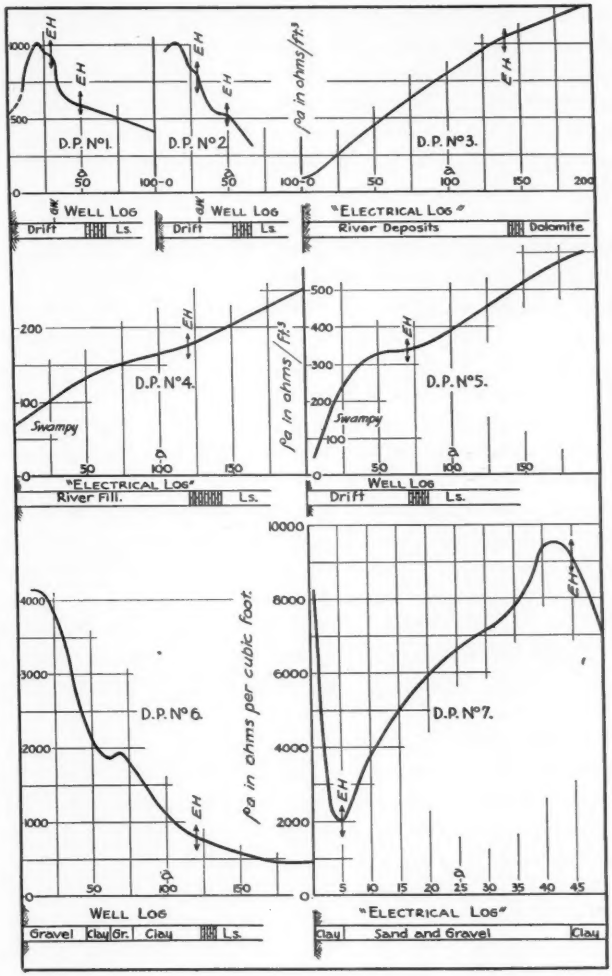

FIG. 3. Typical earth-resistivity depth profiles with well log or electrical log plotted adjacent.

rapid than for Wenner's Method, as only the two inner electrodes are moved. The general characteristics of the two depth profile curves are similar, the electrical horizons (EH) occurring at the same values of "a" for both curves. The slope of the lower part of D. P. No. 2 is different from that of No. 1, which might be accounted for by the fact that heavy rains occurred between the time these curves were obtained.

Depth profiles Nos. 3 and 4 were run on the flats along the Mississippi River south of St. Paul, Minnesota. For both these curves the surface conductivity was good, but for D. P. No. 3 sands and gravel were near the surface while for No. 4 the surface was covered with peat. A well near D. P. No. 4 drilled to a depth of 100 feet found no bedrock. The earth-resistivity data indicates an electrical horizon at "a" = 120 feet, the probable depth to bedrock. For D. P. No. 3, bedrock would be expected at 140 feet in line with the fact that No. 3 was nearer the center of the aggraded gorge.

Depth profile No. 5 was also run over peaty terraine. The initial resistance values are low, as ground water was within two feet of the surface. Below the peat, drift is indicated by the increase in resistance. As the value of "a" approaches 70 feet, the curve flattens, a characteristic often noticed when approaching the Platteville limestone. Beyond "a" = 70 the apparent resistance increases. Bedrock is picked at this depth, checking a well near the profile.

The depth profile shown as D. P. No. 6 was run over a gravel deposit west of Minneapolis. The initial resistances are high, as would be expected over gravels. These high values drop rapidly on approach to the water table. The depth of the water table calculated by Tagg's formulas [24] appeared to be 35 feet. At Station "a" = 70 feet, a resistivity high may mean more gravel, of little thickness however, as the curve continues with the same slope as before. At "a" = 120 feet the curve flattens out, definitely indicating the presence of bedrock. The actual bedrock depth, taken from the driller's log of a well near by, is 123 feet.

In connection with a gravel survey conducted for the Minne-

[24] Tagg, G. T.: *Op. cit.*

sota State Highway department, depth-profiles were used to determine the vertical section over the deposit. D. P. No. 7 is one of these curves run on the top of a glacial kame. From this depth profile the depth of the weathered stripping as well as the total thickness of the deposit can be obtained. The stripping depth was checked by a test hole, and the base of the sand and gravel section was the same as the height of the kame above the surrounding country. It has been found that depth profiles and especially step-traverses offer a valuable technique for prospecting and exploring sand and gravel deposits. A paper on this phase of earth-resistivity surveying will be published later by Mr. Wilcox.

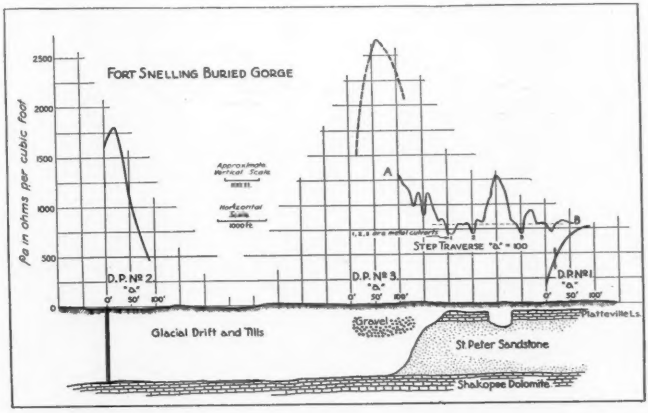

FIG. 4. Depth profile and step-traverse profile and interpretation of the Fort Snelling Buried Gorge.

Fort Snelling Buried Gorge.—South and west of Fort Snelling a wide gorge is known to cross the present Minnesota River valley. This particular area was selected as a good one to demonstrate the practical usefulness of the earth-resistivity method for solving a problem in geology.

Not far from where D. P. No. 1, Fig. 4, was run, the Platteville outcrops in a ditch. This profile is therefore typical of one

where the water table is near the surface, the water being perched on the impervious lime rock. The depth of the Platteville at D. P. No. 1 is less than 20 feet.

Near depth profile No. 2, a well entered bedrock at a depth of 285 feet. The profile at this location definitely indicates the existence of the gorge. The high resistance values are caused by relatively dry drift materials (sands and gravels). As the electrode spacing increased beyond "a" = 20 feet, the apparent resistance decreased. If the profile had been extended to 300 feet an up-turning of the resistivity curve would have indicated the Shakopee Dolomite.

Somewhere between depth profiles No. 1 and No. 2 the edge of the gorge must be found.

Profile No. 3 is a random point. Only three observations were taken, at "a" = 30, 60 and 90 feet. This profile definitely resembles No. 2 and, therefore, must still be over the gorge.

In order to delimit more closely the edge of the gorge, a step-traverse was run from D. P. No. 3 to D. P. No. 1. The electrode interval used was 100 feet. The resulting curve is not smooth because of variations in the drift material; the two "highs," however, are quite convincing. The one between the points marked 2 and 3 can best be explained either as a minor gorge one side of an island or else a tributary to the main gorge. At the points 1, 2, and 3 metal culverts were crossed, the culverts lying in the direction of the traverse. If these lows were corrected, the apparent resistance of the Platteville would be uniformly 800 ohms per cubic foot. Further work should be done, of course, before the complete geologic section can be reconstructed from the electrical data. Another traverse of longer electrode separation would definitely eliminate the remote possibility that the second high was due to a streak of gravel or to a spot of exceptional drainage in the Platteville limestone, assuming it to be present there.

The Westcott Buried Gorge.—The existence of a buried gorge shunting the big loop of the Mississippi River where it passes through St. Paul has been known for many years. Dr. Winchell

pointed it out in 1888.[25] In this general area numerous well logs record the fact that the Platteville and St. Peter formations are missing and that a buried gorge must exist.

Because of the rough terminal-moraine topography and the linear distance to be surveyed to get the desired information, it was decided to use depth profiles rather than a step traverse. The actual points selected for the electrical soundings were made on the basis of accessibility and location with respect to the probable edge of the gorge.

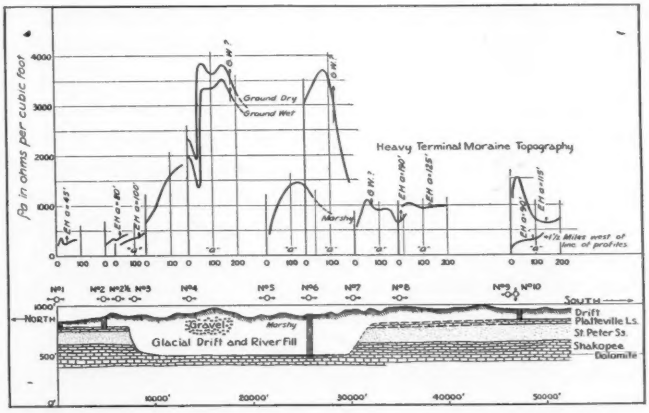

FIG. 5. Depth profiles of the Westcott Buried Gorge.

Depth profile No. 1, Fig. 5, was run for control near a well at the Riverview golf course. At this point the Platteville was buried under 45 feet of drift. At D. P. No. 2 the top of the Platteville appears to be 80 feet deep. A well drilled at this point two months after the electrical survey was taken found the Platteville at 73 feet. Depth profile No. 3 is definitely in the gorge. This curve would turn down and then increase again if it were extended so as to include the Shakopee Dolomite in the electrical section. To do this "*a*" would have to be extended

[25] Winchell, N. H.: Final Report of the Minnesota Geological and Natural History Survey, vol. 2, p. 315, 1888.

considerably beyond 200 feet, its present maximum value. The up-turning of the resistance values as "*a*" is increased is one criterion for a bedrock electrical horizon (*EH*). Another criterion for the presence of the gorge is the high total resistance values of the depth profiles. One would expect high apparent resistance over the gorge because of the thick section of relatively dry glacial tills that now fills it.

D. P. No. 2½ is a fill-in station to locate more definitely the edge of the gorge. A step-traverse between No. 2½ and No. 3, as at the Fort Snelling gorge, would closely delimit the north edge of the buried valley. It is interesting to note that the respective differences in the *EH* values for depth profiles No. 1, 2 and 2½ are about their relative differences in elevation.

Depth profile No. 4 was run twice, once when the surface was very dry, and again after a series of heavy rains. Note the similarity of the two curves and the fact that they approach each other as the value of "*a*" is increased, which was to be expected, as the surface conditions become quite superficial for large electrode separations. The high apparent resistance values for D. P. No. 4 and No. 6 indicate sand and gravel deposits.

The resistance values for depth profile No. 5 are of much smaller absolute magnitude, but of the same general character. Here the surface conductivity was good, as muck was found in a ditch where the profile was run. Depth profile No. 6 was run after a heavy rain.

Profile No. 7 is not definite as to its interpretation. The undulatory nature of the curve suggests heterogeneous drift. The electrical horizon at "*a*" = 190 feet may represent bedrock; however, it would be unsafe to pick this point unless the profile were extended. Profile No. 8 appears to indicate bedrock at 130 feet, and here again the interpretations are not clear, and before definite results could be expected more profiles would have to be taken. Because of the extremely rough topography it was impossible to select suitable sites for the profiles at the south end of the gorge. The wells along the line from profile No. 6 to No. 9 show no bedrock, the deepest one being 150 feet.

At a point one and one-half miles west of depth profile No. 9, Platteville Limestone occurs in a well at a depth of 80 feet. Profile No. 10 was run against this well for a control point. The *EH* on the curve is in good agreement with the well log. This curve resembles D. P. No. 1, 2 and 2½. Profile No. 9 appears also to be definitely over the Platteville. The relative difference in the *EH* values of "*a*" for these two curves is just the difference of their respective elevations.

This line of depth profiles was run in three and one-half days by the authors. The data are far from complete, but enough to indicate the possibilities of the earth resistivity method when applied to geologic problems.

CONCLUSIONS.

The following statements seem warranted by the experience gained in the work described above.

1. It is possible to determine with a fair degree of accuracy the depth of glacial drift overlying the bed rock surface.

2. The curves indicate rather definitely the character of the glacial drift; *i.e.*, whether it is mainly sand and gravel or clay.

3. The resistivity method can be used with a relatively inexpensive machine; the operation is simple and rapid, so that the cost of geologic information obtained in this way is not excessive for general purposes. For specific projects the method could be used at a considerable saving over blind drilling.

RELATION OF CRITICAL AND SUPERCRITICAL PHENOMENA OF SOLUTIONS TO GEOLOGIC PROCESSES.

EARL INGERSON.

INTRODUCTION.

How is gaseous transfer of material from magmas effected? No one well-informed on the subject now doubts that considerable amounts of material are transported by gases, but there is not complete agreement concerning the mechanism of transfer.

There are two entirely separate and distinct modes of transfer possible. These are:

1. By means of volatile compounds; *i.e.*, compounds that are gases under the conditions of temperature and pressure that obtain.

2. By means of gaseous solutions of *non-volatile* compounds in volatile solvents.

Agreement is not perfect among geologists on the condition of the residual solutions at an advanced period in the crystallization of a magma. Niggli (24) [1] would have the entire residual magma above its critical temperature during the pneumatolytic stage. Fenner thinks that this point is probably never reached, and with respect to these residual magmas he says (9, p. 72):

> There is no certainty that such a stage is ever reached. On the contrary, it is possible that at all stages during the cooling of a magma the residual liquid holds enough dissolved material so that the critical temperature appropriate to it always lies above that prevailing at the time.

Morey (21) also believes that critical phenomena have no meaning in the magma. Ross (25, p. 879) expresses a similar opinion. This view is probably correct for the magma, but critical phenomena may be very important in determining the behavior of emanations that have left the magma.

[1] Numbers in parentheses refer to Bibliography at end of paper.

Fenner ascribes important effects to pneumatolytic action, both within the early-formed igneous rock and in the country rock (9, p. 105). He recognizes the possibility that water in the supercritical condition may carry non-volatile material in solution (p. 88). However, as there is little experimental evidence on this point, he prefers to ascribe a much more important rôle to volatile compounds, such as the halides.

Concerning this point Lindgren says (20, p. 709):

> Many writers, like Leith and Harder, Calkins, and Goldschmidt, believe that the heavy metals were introduced as chlorides, fluorides, etc.; and this view is very likely correct. The mode of transfer of silica from the intrusive to the limestone is a problem not fully solved.

These ideas seem to express the consensus of opinion of geologists regarding the gaseous transfer of material. In almost any treatise on contact metamorphism (formation of pyrometasomatic deposits) the author may be found groping for appropriate compounds that are sufficiently volatile to have left the magma as gases. The present tendency seems to be to emphasize the importance of these volatile compounds and to minimize, or eliminate, the possibility that solvents above their critical temperatures may have been effective in carrying non-volatile material to contact zones.

Aim of the Paper.—It is not the writer's object to attempt to minimize the importance of volatile compounds in gaseous transfer—doubtless they are very important in most cases. The object is to emphasize the possibility that solutions above their critical temperatures may also effect a considerable transfer of material, and to call attention to experimental work that has a bearing on the question.

Volatile compounds can account for the transfer of many of the elements from magmas. Vogt (29, p. 217) mentions several that have critical temperatures no greater than that of water. Some of these are:

Compound.	*Tc.*
SiF_4	—1.5° C.
$SiCl_4$	230
$SnCl_4$	319
$TiCl_4$	358
$AlCl_3$	356
$AsCl_3$	356

These and other compounds whose critical temperatures lie below the temperature of formation of pyrometasomatic deposits can account for much of the material in such deposits. However, there are other similar compounds whose aid has often been invoked to explain the transfer of other elements whose critical temperatures lie far above the maximum temperatures attained in contact zones. A few of these are:

Compound.	*Tc.*[a]
[b] $FeCl_2$	1450° C.
[b] $AgCl$	2460
[c] Cu_2Cl_2	2185
[b] $PbCl_2$	1560
[c] $CaCl_2$	2500

Some of these compounds have hardly any measurable vapor pressures at the temperatures of formation of pyrometasomatic deposits. Hence, they must be considered non-volatile compounds at such temperatures, especially under the pressures prevailing during the formation of contact deposits.

Therefore, if such compounds have been effective in gaseous transfer they cannot have been present as gases themselves, but must have been *dissolved* in more volatile compounds that would be true gases under the prevailing conditions.

Definitions.—Before proceeding further it may be advisable to define certain terms so that there will be no misunderstanding concerning their use in the following discussion. Critical temperature is that temperature above which a gas cannot be liquefied by pressure alone. The pressure under which a substance may exist as a gas in equilibrium with the liquid at the critical temperature is the critical pressure. A solution may be defined as any

[a] Estimated from boiling point. [b] (30, pp. 130–131.) [c] (9, pp. 97, 84.)

homogeneous mixture the composition of which can undergo continuous variation within certain limits. The term *gaseous solution* is used in this paper to indicate a solution of a non-volatile solute in a gas above its critical temperature, and not a mixture of gases or vapors. A true gas must be above its critical temperature; a substance in gaseous form below its critical temperature is a vapor. The distinction between a confined gaseous solution and an escaping gaseous solution is an important one. *A confined gaseous solution* moves through the pore space and channels in the rock so slowly that there is no appreciable relief of pressure due to motion away from its source. The maximum pressure is determined by the weight of the overlying rock plus the strength of the crust. An *escaping gaseous solution* has access to the earth's surface and the pressure decreases rapidly to one atmosphere.

EXPERIMENTAL WORK.

When a solution passes through its critical point into the supercritical state does it remain a solution? Or does it become a mixture of gases, with the separation of any solute or solutes that are not volatile at that temperature? Is it not possible for water above its critical temperature to carry non-volatile compounds in solution just as hot water may carry non-liquid compounds in solution below that temperature? And could not considerable amounts of material be deposited directly from such gaseous solutions?

Considerable actual experimental work that has a bearing on the subject has been performed by physicists and physical chemists. The first work on critical and supercritical phenomena of solutions was done by Hannay and Hogarth in 1879 (12).

Their experiments were performed with organic solvents, such as ethyl alcohol and ether, and inorganic solutes whose melting points lie above the critical temperatures of the solvents. They found that in some cases the solutions passed through the critical point with no apparent change. Hannay and Hogarth say (12, p. 325): "We are only able to state that the substance does not show any appreciable change at the critical point of its solvent.

Such was the case with anhydrous chloride of cobalt in absolute alcohol." The critical temperature of alcohol is about 238° C. and the vapor pressure of $CoCl_2$ reaches 760 mm. at 1049° C.

In other cases they found that a part of the solute was deposited just before the critical point was reached (probably due to decreasing density of the solvent), but was re-dissolved after the critical point had been passed. They say further (12, p. 325): "Thus, ethyl alcohol, or ether, deposits ferric chloride from solution just below the critical point, but re-dissolves it in the gas, when it has been raised 8° or 10° above that temperature." The boiling point of ferric chloride is 315° C.

The solids remained in solution as far above the critical temperature as the experiments were carried: ". . . even when the temperature was raised 130° above the critical temperature, and the gas was considerably expanded."

Concluding their preliminary note, Hannay and Hogarth say (12, p. 236):

> We have, then, the phenomenon of a solid with no measurable gaseous pressure dissolving in a gas, and not being affected by the passage of its menstruum through the critical point to the liquid state, showing it to be a true case of gaseous solution of a solid.

Hannay believes that a gas must be above its critical temperature before it can act as a solvent for solids, and points out that a gas below its critical temperature is not a true gas, but a vapor. He states (16, pp. 412–413):

> Let a liquid be colored by some non-volatile colored solid dissolved in it, and let it be heated under pressure, the liquid will remain colored while the vapor will be quite colorless, and will remain so up to the critical point.
>
> Now let the fluid be raised above its critical point, all the internal space will be colored, showing that the gas dissolves the solid while the vapor does not . . . gases are solvents of solids; vapors are not.

In this work no measurements of pressure were made. It seems logical that there must be a confining pressure in the order of magnitude of the critical pressure before a gas can act as a solvent for a non-volatile solute. With a given solute and gaseous solvent there are two factors that determine the solubility; (1)

the density of the solvent and (2) its temperature. Or, as Hannay expresses it—"molecular closeness and *vis viva*" (15, p. 486). Density is dependent upon pressure, so there must be some minimum pressure below which a given solvent at a given temperature will not dissolve an appreciable amount of a given non-volatile solute. Just how small may this pressure be, say for water at 575° C. carrying the salts commonly found in hot springs? This question cannot be answered at the present time. Hannay makes the following generalized statement about pressure, without giving any numerical data (15, p. 486): "The gas must have a certain density before it will act as a solvent, and when its volume is increased more than twice its liquid volume its solvent action is almost destroyed."

He found that the solubility of a non-volatile solute in a gas increases with increasing temperature, and says concerning this point (15, p. 486): "Retaining the volume the same, the higher the temperature the greater the solvent power."

Evidently these factors are interdependent; the higher the temperature, the lower the minimum pressure required for a gas to act as a solvent. No numerical data have been obtained in this field and experimental data concerning the solubility of various solutes and combinations of solutes in water at different supercritical temperatures and at various pressures would be very interesting and of considerable importance in the interpretation of certain geologic phenomena.

Centnerszwer and his co-workers (3, 4, 5, 6, 7) have performed a number of researches on critical phenomena of solutions, but apparently have not concerned themselves with supercritical phenomena. Of their many interesting conclusions only a few need concern us here. Centnerszwer says (3, p. 500): "The material in question raises the critical temperature of the solvent considerably." . . . "This raising is proportional to the concentration of the dissolved material."

Centnerszwer and Pakalneet concluded further that (5, p. 313): "The raising of the critical pressure is proportional to the concentration of the solution."

Schröer corroborated these conclusions in 1929 (27, p. 390). These facts of the elevation of the critical temperature and pressure of a solvent by dissolved solute are frequently mentioned in geologic literature, but rarely is a reference given in their support.

This raising of the critical constants, however, is due to non-volatile dissolved material. Mixtures of two volatile substances may show a minimum critical temperature that is below the critical temperature of either of the pure substances. This is an important point and should be kept in mind in any discussion of "the volatile components in a magma," and more particularly after they have left the magma. It is commonly assumed that any impurity will elevate the critical temperature of water, but this is not an unavoidable conclusion. Conceivably, some of the other volatile components from a magma might lower the critical temperature of water and thus at least partially compensate for the elevation brought about by the non-volatile solutes. Centnerszwer and Zoppi (4, pp. 701, 705) mention ether and methyl alcohol as a system showing such a minimum critical temperature. They state: "Die kritische Linie des Systems Äther-Methyl-alkohol, weist ein Minimum auf."

Centnerszwer and his co-workers used organic solvents. Their numerical data on the critical temperatures need not concern us here, because more pertinent to the question at hand are the results of Schröer, which were obtained with water and non-volatile halides. He made a number of determinations with solutions of the alkali halides which ranged in concentration up to about 5 per cent. Some of the most concentrated solutions gave the following results (26):

Concentration Mol/liter.	Solute.	Elevation of Tc.
.810	NaCl	36.5° C.
.684	KCl	38
.342	KI	25.15

Schröer gives no data on the elevation of the critical pressure of water. Some are given, however, in earlier works for organic solvents. Centnerszwer and Pakalneet (5, pp. 307–308) give

the following data for solutions of triphenylmethane in ethyl ether:

Concentration Per cent. per 100 g.	Tc.	Pc.
0	194.3° C.	36.2 Atm.
2	200.9	38.4
4	206.8	40.8
8	218.9	45.1
16	240.5	54.7

Van Nieuwenburg and Smit (28, p. R-224) have shown that silica is appreciably soluble in water at temperatures from 365 to 410° C. and under pressures of 200 to 350 atmospheres.

APPLICATIONS TO GEOLOGIC PROCESSES.

The experimental work cited above is rarely mentioned directly in geologic literature. Apparently the first mention of it was made by Boeke in 1915 (1). In the second edition of this work by Eitel in 1923 (2, pp. 347–348) all of the work accomplished up to that time is mentioned. However, no work had then been done using water as a solvent. Also no extensive applications to geology are indicated, since the work is essentially a summary of physical-chemical data that are applicable to petrology.

Knopf seems to have made the first definite geologic application of the transfer of non-volatile material in gaseous solution. In discussing the origin of the ore deposits of the Yerington District, Nevada, he refers to the work of Hannay and Hogarth. He indicates clearly that aqueous solutions in the supercritical condition may have been effective in the formation of the deposits (18, p. 45).

It is fairly well established now that water is the overwhelmingly predominant constituent (98% ±) of most volcanic exhalations. Presumably this is true also of the material given off by crystallizing plutonic magmas. Fenner (9) believes that these solutions leaving the magma are dilute. If these conclusions are justified, then the data presented above may be applied directly to certain geologic processes.

1. It has been shown experimentally that non-volatile solids may be dissolved in gases. This provides a mechanism whereby elements can leave a magma without the necessity of existing therein as volatile compounds. Even silica might leave the magma in this manner, according to Greig, Merwin, and Shepherd, who state (11, p. 71):

> . . . at high temperatures the presence of a small pressure of the "volatiles" from rocks, or of water, provides whatever is necessary to bring about the transfer of silica through the vapor phase at an important rate, while without these volatile materials, under conditions identical in every other respect, the rate of transfer, if real, is enormously slower, being indeed inappreciable by the rather sensitive methods used.

These experiments were carried out at temperatures from 600° C. to 1200° C. and under only a few atmospheres pressure. The authors do not attempt to explain just how the transfer is effected, saying that the rôle of the water is unknown. Is it possible that this is a case of solution of silica in water above its critical temperature? This does not seem likely in view of the comparatively slight pressures used. At any rate, it is a well-defined case of the transport of silica by water above its critical temperature without calling for the aid of a halogen. If the amount of transfer increases, or at least does not decrease, with increasing amount and density of the solvent, it is evident that water above its critical temperature could carry silica from magmas without the necessity of having volatile compounds of silicon involved. The work of Van Nieuwenburg and Smit (28) supports this conclusion.

2. The transport of non-volatile material in gaseous solution provides a means of transferring abundant non-volatile material to pyrometasomatic ore deposits. It may also aid in the explanation of some of the features of such deposits. Fenner points out that deposits formed at depth by gases would differ in no way from those formed by hydrothermal solutions. He states (9, pp. 91–92): ". . . but there is no reason to suppose that, at deeper levels, dense, solid rocks may not result from metasomatic reactions by gases as well as from the hydrothermal solutions or from orthomagmatic crystallization."

This conclusion is probably valid in most cases, but there may be, in certain instances, ways of distinguishing between deposits formed by gaseous solutions and those formed by hydrothermal solutions. The deposits formed by gases would be more likely to contain elements that form readily volatile compounds than would hydrothermal or orthomagmatic deposits. Minerals whose temperatures of formation lie above the critical temperature of the mineralizing solution could not have formed from hydrothermal solutions.

The depth of formation may serve in some cases to distinguish the two types of deposits. Hydrothermal deposits may form at any depth, but there must be a minimum depth above which deposits from a confined gaseous solution cannot form. (See 3, below, for the case of *escaping* gaseous solutions.) This depth will be determined by the temperature and concentration of the solution and the "Füllungsgrad."[2] For a gaseous solution of given temperature and concentration there must be a certain minimum pressure or the solution will either (*a*) deposit its dissolved material because of decreasing density, or (*b*) overcome the pressure and break through to the surface. The Füllungsgrad will determine which one of these possibilities takes place.

Assuming that the pressure must be at least as great as the critical pressure for the solution before the gas will have any considerable solvent power, then it is possible to calculate the depth above which deposits from confined gaseous solutions may not form. If the concentration of these solutions is of the same order of magnitude as those with which experiments have been performed, then the value of the critical pressure can be estimated. The increase of the critical pressure for ether is about 32 atmospheres for an .8 $\pm$ mol/liter solution.[3]

If the increase is in the same order of magnitude for water solutions, then the critical pressure for solutions such as the most

[2] This handy term is used by Centnerszwer (3, p. 454) to indicate the ratio of the liquid to the available space.

[3] Calculated from the data given by Centnerszwer and Pakalneet. The molecular weight of triphenylmethane is 244.3 and the specific gravity of ether is .719.

concentrated of those used by Schröer would be 217 (+) + 32 = 250 ± atmospheres. If 2.7 is taken as the average density of the earth's crust, it follows that deposits from such a gaseous solution would be formed only at depths greater than some 3100 feet.

This figure is not inconsistent with the inferred depths of formation of pyrometasomatic deposits. Lindgren (20, pp. 799–800) mentions several such deposits that are thought to have formed at depths of 2,000 to 4,000 feet. He says that Spencer believes the deposits at Cornwall, Pa., to have formed at a depth of about 1,000 feet. If these deposits, or any of the minerals in them, were formed from gaseous solutions, then the formation of these deposits less than 3,000 feet below the earth's surface lends weight to the suggestion made above that superheated gases may carry non-volatile material in true solution at pressures lower than their critical pressure. It would be necessary to assume in this case that there is not enough solution present to fill the available space at the critical density of the solution, because if there were its vapor pressure at or above the critical temperature would overcome any external pressure less than the critical pressure. The strength of the crust must also be taken into consideration. This would tend to increase the pressure so that it might be possible to have the pressure at 2,000 feet greater than the critical pressure. Obviously, it is impossible to evaluate all of these factors at present.

A consideration of the effect of the degree of filling of available space involves difficulties beyond the scope of this brief paper. Centnerszwer (3) insists that the Füllungsgrad exerts an appreciable effect on the critical temperature, and he gives a number of experimentally derived curves in support of this view (pp. 485–492). His conclusion on this point is (3, p. 500): "Die kritische Temperature der Lösung fällt dauernd mit steigenden Füllungsgrad, im Gegensatz zu dem Verhalten reiner Substanzen, wo sie ein Maximum erreicht."

However, Hannay is quite positive that the degree of filling has no influence on the critical temperature (15, pp. 486–487). Con-

cerning pure water Schröer says (26, p. 109): "The appearance of the critical state is independent of the Füllungsgrad in the region investigated (where the value of the Füllungsgrad is 0.28 to 0.39)."

Evidently, exceptional conditions attend the formation of pyrometasomatic deposits like that at Cornwall, Pa., because so few have been reported with such depths of formation. Recent work by Hickok [4] indicates that the depth of formation at Cornwall may have been as great as 5,000 feet. A depth of 2,000 feet seems to be the upper limit for nearly all such deposits. A depth of 2,000 feet in the earth's crust corresponds to about 160 atmospheres pressure. The majority of pyrometasomatic deposits probably, according to Lindgren, form below 1000° C., since siliceous rocks consolidate at temperatures of 500–1100° C. (20, pp. 710–711). It would be interesting to determine how much of various non-volatile solutes water can carry in solution at these temperatures, especially 500–800° C., and 160 atmospheres pressure. These temperatures are, of course, far above the critical temperatures of moderately dilute mineralizing solutions that are under consideration. These temperatures would be about 374° C. (critical temperature of pure water) + 38° C. (maximum elevation of the critical temperature of a water solution found by Schröer) = 412° C.

Primary mineral deposits formed from confined solutions nearer the surface than about 2,000 feet are hydrothermal deposits. Of course, many deposits formed below that depth are also of hydrothermal origin, so that only in especially favorable instances would depth of formation definitely distinguish hydrothermal deposits from those formed from gaseous solutions. However, it is commonly more difficult to estimate the depth of formation than it is to establish the mode of origin by other criteria. When the temperature-pressure relations for the formation of the minerals of such deposits are more fully known, they may furnish a key not only to the state of the solutions that formed the deposits, but also to the depth of formation.

[4] Hickok, W. O.: Econ. Geol., vol. 28, pp. 193–255, 1933.

The above discussion of depth of formation is based on the critical pressure of a dilute aqueous solution. Similar arguments could be applied to the case of volatile compounds other than water. Other compounds are neglected for two reasons: (1) In the first place, water is assumed to be by far the most abundant constituent of the solutions and the critical pressures would not differ greatly from that of water. (2) The critical pressures of the other volatile substances are not known and so cannot be taken into consideration at this time.

3. Conditions are quite different when a gaseous solution has a channelway by which it may come to the earth's surface. Under such circumstances it is possible that near-critical conditions might obtain fairly close to the surface; at least, it is quite conceivable that the conditions of temperature and pressure could be such that the water (as a gas) could carry non-volatile solutes nearly, or quite, to the surface. The dissolved solids would be deposited in and around the channelways as the temperature and pressure decreased.

The well-known volcanic sublimates found in the Valley of Ten Thousand Smokes, around Vesuvius, and in many other places, seem to be formed largely from volatile compounds. They commonly either contain such compounds, or the gases coming from the fumaroles contain elements that would form volatile compounds with the metals present in the deposits. Such deposits may form by a simple decrease in temperature, hydrolysis, reaction with the wall rock, reactions of gases, or by combinations of these or other processes.

The temperature of fumarolic gases is commonly above the critical temperature of dilute solutions, so it is quite possible that gaseous solutions (non-volatile matter dissolved in a solvent above its critical temperature) may contribute some material to these deposits. This view is not urged for sublimates that can be explained by assuming volatile compounds, but it may be the explanation of some so-called "sublimates" such as those of calcium mentioned by Fenner (9, pp. 84–85), who states:

> The very low volatility of calcium chloride is puzzling in view of con-

siderable evidence that indicates that calcium salts have been volatilized and deposited as volcanic sublimates. . . . Maier found no detectable pressure of $CaCl_2$ at 1137.6° C. . . .

However, $CaCl_2$ is readily soluble in water, and these calcium-containing deposits may be explained by assuming that water above its critical temperature carried calcium salts in solution. Under this interpretation they would not be sublimates in the strictest sense of the word, but would be sublimate-like deposits formed from gaseous solution.

Such deposits may form clear up to the surface of the earth from an *escaping* gaseous solution. The conditions are entirely different from those obtaining in the formation of a pyrometasomatic deposit near an intrusive mass, where movement through the pore space is too slow to relieve the pressure. Thus, the fact that escaping gaseous solutions (but more commonly escaping vapors) may deposit material at the earth's surface should not be construed to invalidate the conclusion that pyrometasomatic deposits formed from *confined* gaseous solutions have a minimum depth of formation. Zies (31, pp. 17–21) discusses these open and closed systems.

The difference in the physical properties of these two types of deposits is sufficient evidence of the difference in genetic conditions. It is necessary to keep clearly in mind the distinctions between vapors and true gases; also between gaseous solutions involving non-volatile compounds and mere mixtures of gases.

CONCLUSIONS.

1. All gases above their critical temperatures can probably carry non-volatile solids in solution. Water is the most important solvent from the geologic point of view. It has been shown that water above its critical temperature can carry at least 5 per cent. of its weight of alkali halides and, independent of the alkalis, appreciable amounts of silica.

2. The critical temperature and critical pressure of such solutions of non-volatile solutes exceed those of the pure gas by an amount that is proportional to the concentration of the solution.

The presence of other volatile substances may bring about a decrease of the critical temperature.

3. Gases at supercritical temperatures may dissolve non-volatile solutes even though the pressure is below their critical pressure. Experimental work is needed in this field.

4. The fact that gases may carry non-volatile material in solution proves helpful in explaining some features of

a. Gaseous transfer of non-volatile material from magmas
b. Pyrometasomatic deposits, as to (1) contained substances (even silica), and (2) depth of formation.
c. Volcanic "sublimates" of non-volatile substances.

5. Undoubtedly much gaseous transfer of material is effected by means of volatile compounds, but gaseous solutions are probably important in certain cases.

6. The experimental work is quite definite as far as it goes, but very little work has been done on geologically important substances. Therefore, the applications to geology are mere suggestions because it is impossible to state at present just how far these principles may be applied geologically.

It is hoped that this note will be helpful in starting to remove the "stumbling block" (9, p. 61) of critical phenomena from the path of the geologist.

ACKNOWLEDGMENTS.

The writer is indebted to Professor Knopf not only for the inception of the idea that the transport of non-volatile material in gaseous solution may be geologically important, but also for certain references to experimental work and for various suggestions and criticisms. The writer is also indebted to Professor Bateman for the suggestion that the paper be written up in its present form, for additional references, and for many valuable suggestions and criticisms concerning the manuscript. Thanks are also due to Dr. C. N. Fenner, of the Geophysical Laboratory, for reading a preliminary copy of the manuscript and making helpful comments and suggestions.

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TIME–TEMPERATURE CURVES IN RELATION TO MINERAL ASSOCIATIONS IN COOLING INTRUSIONS.[1]

HANS SCHNEIDERHÖHN.

ACCORDING to definition, a hydrothermal ore body is one formed by aqueous solutions below their critical temperature but above boiling point; that is, in round figures, between 400° and 100° C. The critical temperature of the solvent does not define a sharp boundary, but hydrothermal deposits may grade into pneumatolytic deposits. The entire form and paragenetic characteristics of the two kinds of deposits are so typical and unmistakable, however, that in every true genetic classification they must be regarded as separate groups.

The cooling of the easily volatile constituents to hydrothermal temperatures at first takes place at some distance from the solidifying intrusive rock; but at later stages it occurs progressively nearer to the igneous mass, and after crystallization of the mass, may take place within it. The places and depths of hydrothermal mineralization vary considerably during the magma crystallization. According to Lindgren, the depth varies from 1,000 to 3,000 m. at the time of deposition. Data concerning this were obtained from several American occurrences where the amount of denudation could be calculated. This depth also corresponds well with that of formation of hypabyssal igneous masses. The depth of hydrothermal mineralization may rise to within a few hundred meters of the surface in regions of long and vigorous intrusive activity. On the other hand, it may occur at depths greater than 3 kilometers, chiefly at the beginning of an intrusive cycle, in a region with long periods of intrusive inactivity, during

[1] This paper was written as a contribution to a presentation volume to Professor F. Rinne on his seventieth birthday, March 16, 1933. Unfortunately, his death four days earlier prevented this. Translated from German by A. A. Cawley and the editor.

which the country rock has cooled. The rock temperature at depths of 1,000 to 3,000 m. may vary from 50° to 125° C. Temperature, however, does not depend on depth alone, for even at the depths of hydrothermal mineralization the rock temperatures have been locally raised by the magma. Most hydrothermal deposits are probably formed between 175° and 300° C., although mineralization perhaps begins at 400° and may continue to 100° or less, when minerals are being deposited in only small quantities in the beginning and end stages.

On the assumption that a body of crystallizing magma is cooling at depth, the consequent temperature changes will be considered (1) in the intrusive itself, (2) at the contact, and (3) at various places in the country rock. The important results of these considerations will be applied to a *critical examination of mineral succession in hydrothermal ore bodies.* Other factors that influence mineralization, such as pressure, differences in country rock, texture, structure, jointing, will not be considered. The isogeotherms are not concentric spherical forms, nor do they always remain at the same depth, but each advance of the magma sets them in motion and disturbs their normal arrangement. They bulge out strongly above the intrusion and simultaneously move closer together. With advancing crystallization and cooling, they become flatter inward and more widely spaced. Hence, in the neighborhood of an intrusion, at the time of the intrusion, the temperature gradient in the overlying zones is much steeper than normal.

In consequence of these temperature changes, *lateral and temporal facies changes,*[2] as P. Niggli has appropriately called them, occur. Places in the same fissure at different distances from the intrusive body have, at the same time, different temperatures, and the mineralization shows a "lateral" change. In the course of time, a given point in a fissure passes through different temperatures, i.e. a "temporal" change in mineralization occurs. Both these types of change in hydrothermal mineral associations

[2] "Laterale und temporale Fazieswechsel." "Facies" is here extended from sedimentary to igneous deposits.

combined are what is known by the old German mining expression "Teufenunterschiede" (depth differences).

The derivation of time and temperature curves will now be attempted, to show the temperature changes at different points in the country rock around an intrusion, and in the intrusion itself, during and after crystallization. Reference should be made to the important work of Ingersoll and Zobel.[3] They discuss the course of cooling in a laccolith and the flow of heat in the country rock, but in order to solve the problem they introduce simplifications that do not altogether correspond to natural conditions. The cooling mass is assumed to be spherical, with a radius of 1,000 m., and surrounded by an infinite quantity of a homogeneous country rock with an initial temperature of 0° C. throughout. The thermal data for granite and limestone are used; the latent

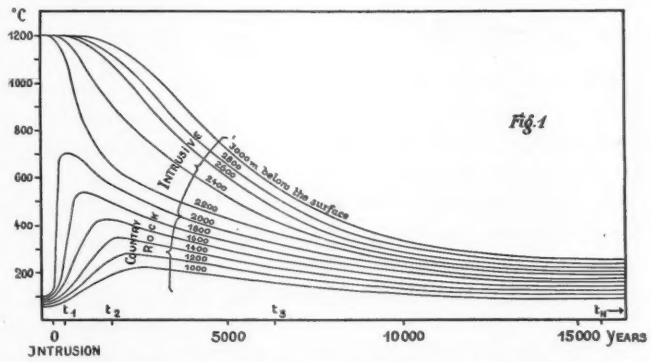


FIG. 1. Temperature-time diagram showing progressive cooling of an intrusive mass and its contact rocks for distances of from 200 to 4200 meters. Reconstructed after Ingersoll and Zobel. Time and temperature, approximate.

heats of crystallization were not considered. In their Diagram 23, page 127, is given a series of curves calculated for 100, 900, 3,600, and 10,000 years after intrusion, which show the fall in

[3] Ingersoll, L. R., and Zobel, O. J.: An Introduction to the Mathematical Theory of Heat Conduction, with engineering and geological applications, p. 171. Boston, 1913.

temperature from the center of the magma mass to a point about 600 m. distant from the contact. From these curves I have derived another diagram, with temperature as ordinates and time as abscissas, each curve corresponding to a certain distance from the magma-center. These curves are plotted at 200 m. intervals, up to 1,000 m. from the contact in country rock (Fig. 1).

To simplify calculations, the following assumptions were made: (1) The initial temperature of the magma was 1200° C. (2) The intrusive body is a cylinder of 500 m. radius and indefinite depth, and has a superimposed hemisphere. (3) The highest point of the intrusive is 2,000 m. below the surface. (4) The country rock is homogeneous and, before the intrusion, had the following temperatures:

1,000 m. below the surface	60° C.
2,000 m. below the surface	90° C.
3,000 m. below the surface	120° C.
4,000 m. below the surface	150° C.

With these additional postulates, the problem is not susceptible of easy solution, but using the data of Ingersoll and Zobel, the modification of the derived curves by these extensions may be investigated and determined.[4]

The cylindrical form of the intrusive body creates a rising flow of heat, whereas the heating effect toward the sides is less pronounced. Also, the initial temperature of the country rock decreases toward the surface; consequently, the heating effect is greater in this direction. Thus, the time-temperature curves for the contact and for more remote points represent a greater heating effect than do the curves derived from the data of Ingersoll and Zobel. The temperature maxima are greater in the new curves. The general rate of cooling is less, and the time of heating of the country rocks is longer. Fig. 1 shows the curves derived as the result of these considerations. No claim is made for their absolute accuracy, since there are too many quantitative factors that cannot be determined. Nevertheless, the general effects of the system are clearly indicated.

With the temperature of the country rock determined, it is now

[4] The advice of Dr. J. Koenigsberger of Freiburg is gratefully acknowledged.

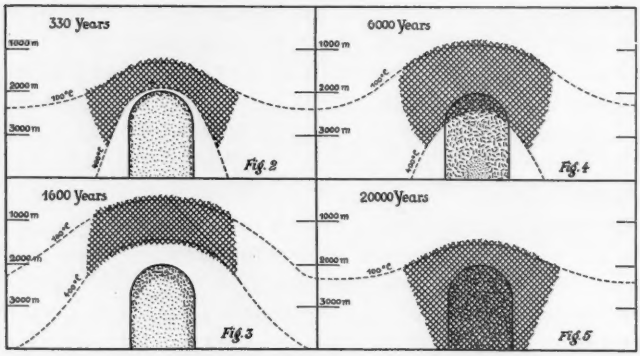
possible to consider the mineralization, since it depends upon the temperature. The temperatures of the mineralizing solutions may be considered negligible, since they are always somewhat higher than that of the country rock. In a recent paper by Boydell [5] this factor is disregarded by assuming a more rapid flow of the mineralizing solutions, but in the present problem it can be neglected by assuming a slow flow.

The cooling curves within the intrusive body (Fig. 1) show a steadily decreasing course, i.e. at every point the temperature decreases with increasing time. Or, in other words, at any given point within the eruptive rock pneumatolytic and hydrothermal minerals separate. On the contrary, the cooling curves in the country rock have a maximum which becomes more pronounced nearer the contact, hence each point passes twice through a given temperature. If minerals were deposited continuously at any point, a facies change from lower to higher, and then again to lower-temperature minerals, would be expected. But actually no mineralization, or only a very little, will occur during the initial rising part of the curve, since at that time the magma is only slightly crystallized and therefore only a small quantity of easily volatile constituents is given off. The greatest quantity is liberated from time t_2 onwards, so that actually a steadily decreasing temperature series from high to low temperature minerals is to be expected. It is possible, however, that in certain cases mineralization can occur where the curve rises, thus yielding younger higher-temperature minerals with older lower-temperature minerals; or the younger higher-temperature minerals may alter or replace the older lower-temperature minerals.

Vertical sections of the Earth's crust, together with the related isogeotherms, appear quite different at different times. In Figs. 2–5, such sections, obtained from the data of Fig. 1, are shown for times of 330, 1,600, 6,000, and 20,000 $\pm$ years after the beginning of the intrusion. These times were chosen because they indicate particularly well the thermal changes during cooling.

[5] Boydell, H. C.: Temperature of Formation of an Epithermal Ore Deposit. Bull. Inst. Min. and Met. London, 1932.

Only the extreme isotherms of hydrothermal mineralization (i.e. 100° and 400° C.) are given, and they inclose the space in which, at the particular time, and with the above-mentioned limitation, hydrothermal mineralization can generally occur. Further, the



FIGS. 2–5. Temperature-time curves during intrusion. Isotherms are given only for hydrothermal formation of minerals at 100° and 400°. Cross-hatched area = region most favorable for ore formation; dashed = crystallized intrusive; stippled = fluid parts. Time values, approximate.

zone favorable to hydrothermal mineralization is, as shown by mining experience, a dome-shaped body with the under side narrower than the upper, over the intrusive body. This is clearly shown in the illustrations, which also indicate how the places of hydrothermal mineralization continually change with change of time. If all these zones are superimposed one above another in a single diagram, the entire trend of mineralization is shown (Fig. 6). Because of the separate times chosen, these fields appear sharply bounded in the diagram, but in reality they merge imperceptibly. However, the typical characteristics are not lost as a result of this. The depositions at the four times indicated in Figs. 2–5 are shown in Table I as groups I–IV. They display the temporal facies change. Each consists of high, intermediate, and low temperature generations. The fields *A–F* of Fig. 6

show the lateral facies change, and so the zonal distribution. The space-time distribution is shown in Table I.

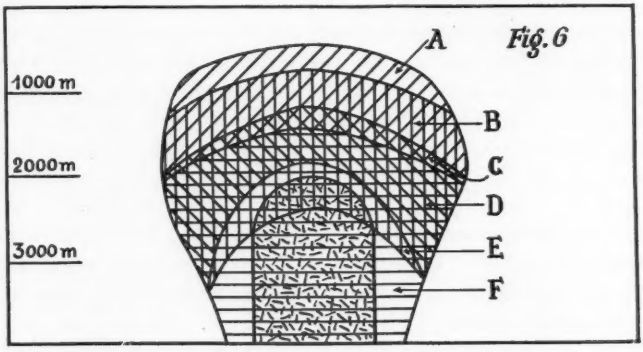


FIG. 6. Composite of Figs. 2–5, to show complete trend of metallization near an intrusive mass after cooling; it shows zonal succession—lateral facies changes in different places and also temporal facies changes in the same place.

TABLE I

Temporal facies change:		I.	II.	III.	IV.
Formation time in years:		0–330	330–1,600	1,600–6,000	6,000–20,000
Ore-Zones in order of depth (lateral facies change)	*A*	—	Low temperature	—	—
	B	—	Intermediate temperature	Low temperature	—
	C	Low temperature	High temperature	Intermediate temperature	—
	D	Intermediate temperature High temperature	—	Intermediate temperature	Low temperature
	E	—	—	High temperature	Intermediate temperature
	F	—	—	—	High temperature

Several relations of the greatest interest result from this tabulation. The zonal distribution of the ore around the intrusion is clearly shown. Quite different depositions follow one another

from the deepest zone *F* in the intrusive body, over the contact zone *E*, to the outermost apomagmatic zone *A*. Further, the temporal parageneses I–IV are generally different, as various distillation changes tend to occur, due to progressive cooling and gassing of the magma. In many of the zones *A–F*, several generations are deposited, and this is what is known as "recurrence," an exceedingly common phenomenon in ore deposits The general rule is that younger generations are of lower temperature than the older. The tabulation of Table I shows that this is so in the majority of cases in the zones *B*, *D* and *E*. But there is an intermediate zone *C* where, after a low temperature generation of group I, a high temperature generation of group II follows, and then again an intermediate temperature generation of group III occurs. Such cases are actually known. In 1928, Berg [6] studied the phenomenon and called it, "rejuvenation." He believes that "rejuvenations" can be explained only by later magmatic activity in old igneous massifs. This possibility is not disputed, but the arguments presented above indicate that in the normal course of cooling of a single magma, there are zones where the isogeotherms rise and fall, and in the same place there is the possibility of first, low temperature, then high temperature, and again, low temperature mineralization.

Rejuvenation need not, therefore, belong to another cycle, but can be at certain places, and under certain conditions, not only a normal, but even a necessary phenomenon in the same cycle. Whether rejuvenation is explained, as by Berg, by a new cycle, or whether it depends upon higher temperature depositions in the same cycle, must be investigated for individual cases. The latter explanation should always be accepted when the depositions of different generations show a close consanguinity. The best example of such a rejuvenation thus far described [7] is that of the hematite-pyrrhotite paragenesis of the Siegerland siderite veins, with high temperature, but younger hydrothermal generations.

[6] Berg, G.: Über den Begriff der Rejuvenation und seine Bedeutung für die Beurteilung von Mineralparagenesen. Zeit. f. prakt. Geol., 36, pp. 17–19, 1928.

[7] Schneiderhöhn, H.: Vorläufige Mitteilung über pyrometamorphe Paragenesen in den Siegerländer Spateisensteingängen. Zeit. f. Krist., 58, 310–329, 1923.

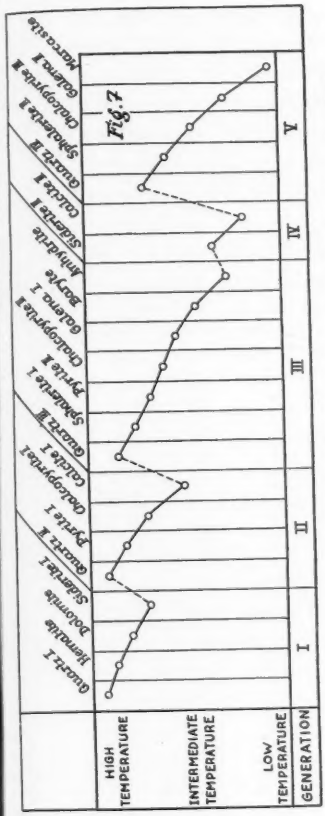


FIG. 7. Age sequence in the lead-zinc veins in Oberharz, showing normal succession and several normal recurrences. After G. Hegemann, 1931.

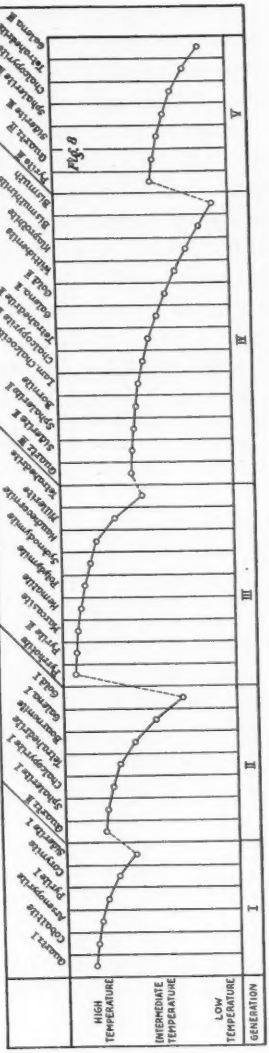


FIG. 8. Age sequence in the Siegerland siderite veins, showing normal succession, several recurrences, and a rejuvenation. After Schneiderhöhn, 1923, and M. Hüttenhain, 1932.

Careful microscopic examination of the veins [8] shows that in the mineral successions at all places the same metals are represented, but in different concentrations and, in consequence, partly in the form of different minerals (Fig. 8).

Finally, Figs. 7 and 8 show satisfactory agreement between the results of such theoretical arguments and the actual occurrences disclosed in carefully examined ore bodies. The paragenesis was determined by careful microscopic examination of much material.[9] Absolute temperatures cannot yet, unfortunately, be stated, but the general character of the high, intermediate, and low temperature mineralization is definitely known.

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[8] Hüttenhain, D. M.: Die Elemente Gold und Wismut als Gangkomponenten der Siegerländer Spateisensteingänge. Min.-petr. Mitt., 42, pp. 285–317, 1932.

[9] Hegemann, G.: Der Vorgang der Spaltenfüllung auf den Oberharzer Bleizinklagerstätten. Neue. Jahrb. f. Min. A. BB., 62, 423–476, 1931.

THE LLALLAGUA–UNCIA TIN DEPOSIT.

VICTOR SAMOYLOFF.

INTRODUCTION.

THIS paper presents the result of field study undertaken by the writer in 1930 and 1931, and subsequent laboratory study at the University of Lyublyana, Yugoslavia. The writer desires to express his gratitude to Prof. V. V. Nikitin of that institution for aid in the laboratory work and in the preparation of this paper. Thanks are also due to Dr. G. B. Barbour for reading the manuscript, and to Mr. J. C. Pickering, General Manager of the Patiño Mines and Enterprises, for permission to publish this article.

GENERAL GEOLOGY.

Llallagua-Uncia, the world's largest deposit of tin ore, is confined to the Cerro de Llallagua, situated in the central part of the Cordillera Real or Bolivian Eastern Range of the Andes, between latitudes 18° and 19° S. The greater part of this range is made up of Paleozoic sediments, consisting of graywacke, sandstone or quartzite, and shale (which, according to d'Orbigny, Steinmann, Kozlowski, and others, are Silurian and Devonian), overlain by the red brackish-water sandstones, limestone, and other strata, of Steinmann's "Puka" formation. Paleontologic evidence indicates that the lower horizons of the latter are probably early Triassic or Jurassic, and the upper, Cretaceous.

The building of the Cordillera Real took place in two stages, the strata being first folded and uplifted with granitic intrusion at the end of the Paleozoic, and then folded again in Tertiary time, with accompanying spasmodic intrusive and extrusive igneous activity, which probably continued into the Pliocene. This epoch was doubtless the period of tin mineralization, giving rise to the Llallagua-Uncia deposit, which has not only economic importance but scientific interest.

The marked folding and faulting of the sediments which preceded the intrusion and ore deposition developed a large overturned anticline with a smaller syncline on the northeast. Comparatively sharp bending produced fissures in the crest of the principal flexure, and magma was forced up at the place where erosion had weakened the anticline down to the Paleozoic Cancañiri graywacke to form a single small body with few apophyses. In plan, this intrusion is roughly elliptical, with dimensions of about 2,000 and 1,800 meters. It contracts with depth and has been followed downward for 778 feet. This form led to the conclusion that the intrusive mass is a small stock; but other evidence indicates solidification in a volcanic neck at or near the surface.

The Country Rock.

The Llallagua-Uncia intrusive is of light and dark shades of gray, due to the presence of tourmaline or volcanic ash, but weathering or strong silicification has changed this color locally to almost white. The texture varies considerably; in general, it is distinctly porphyritic, with a felsitic groundmass carrying phenocrysts of quartz, and less commonly of muscovite and altered feldspar. Also, it ranges from a vesicular, almost spongy, rock to a dense lithic hornstone.

Thin sections show that the dominant phenocryst is quartz. (Fig. 1.) Minute needles of dark-colored minerals are common in them, and there is a remarkable development of microscopic prisms of tourmaline and cassiterite in secondary enlargements of the quartz crystals, oriented along the crystallographic directions of the original phenocrysts. Quartz is the only unaltered original constituent.

The altered rock contains no feldspar. Its place is taken by tourmaline-quartz aggregates, by kaolinite, or rarely, by chlorite. Elsewhere, sharp outlines of replaced feldspars occur as large, tabular, pinkish-white, kaolinized pseudomorphs, twinned according to the Carlsbad law, probably after orthoclase. Biotite, and probably hornblende, have also been altered to tourmaline or tourmaline-quartz aggregates, or to muscovite. The ground-

mass has been replaced by quartz and tourmaline with a resulting fine-grained or felsitic texture.

The post-intrusive effects are limited to the intrusive rock and the sandstone, the graywacke being only slightly changed by the contact action into a compact hornfels. This may be explained by the comparatively low temperature of the igneous body at the time of intrusion, and by the high porosity of the eruptive rock.

Tourmalinization.—Tourmalinization, a characteristic feature of tin deposits, has produced the most marked effect upon the Llallagua-Uncia rocks, indicating abundance and intense action of mineralizers. This alteration is closely related to the deposition of cassiterite, an association long recognized,[1] and considered by some writers [2] to be due to the part played by boron in the formation of cassiterite. For a long time it was generally thought that the Bolivian tin deposits were unique in that they contained no tourmaline. This was especially emphasized by Stelzner.[3] The presence of tourmaline near tin deposits in Bolivia was observed first by d'Orbigny [4] almost a hundred years ago, but close association of tourmaline with cassiterite in the Bolivian deposits was recognized by Forbes,[5] and confirmed by Bonney,[6] Pearce,[7] Spencer,[8] and Rumbold.[9]

The reason why this mineral had been previously overlooked may be ascribed to its unusual occurrence. The Llallagua-Uncia deposit however, is, without doubt one of the finest examples of

[1] McAlister, D. A.: Tin and Tourmaline. Quart. Jour. Geol. Soc., vol. 59, p. 53, 1903.

[2] Ferguson, H. G., and Bateman, A. M.: Geologic Features of Tin Deposits. Econ. Geol., vol. 7, p. 251, 1912.

[3] Stelzner, A. W.: Die Silber-Zinnerzlagerstätten Bolivias. Zeit. Deutsch. Geol. Ges., vol. 49, p. 116, 1897.

[4] D'Orbigny, A.: Voyage dans L'Amerique Meridional, vol. 3, pt. 2, p. 142, Paris, 1846.

[5] Forbes, D.: Phil. Mag., vol. 29, p. 137, 1865.

[6] Bonney, G. T.: The Bolivian Andes, p. 372. M. Convay, 1901.

[7] Pearce, R.: Trans. Roy. Geol. Soc. of Cornwall, vol. 13, p. 105, 1906.

[8] Spencer, L. J.: Miner. Mag., vol. 14, p. 333, 1907.

[9] Rumbold, W. R.: The Origin of the Bolivian Tin Deposits. Econ. Geol., vol. 4, pp. 321–364, 1909.

tourmalinization. Its most interesting occurrence is in nodular pseudomorphs after feldspar and biotite. On exposure, these are more resistant to weathering than the inclosing groundmass, imparting a striking appearance to the rock (Fig. 2). Some of the tourmaline nodules show the distinct prismatic form of feldspar crystals, which serves as a clue to their origin.

Microscopic study shows that these nodules consist essentially of quartz and tourmaline, in varying relative proportions. Remnants of feldspar, represented by aggregates of kaolinite, are commonly present to prove the development of tourmaline from feldspar. The nodules are holocrystalline (Fig. 3); idiomorphic forms of tourmaline are not clearly developed, and the quartz always fills the interstices between tourmaline grains, clearly indicating the order of formation. Other sections show tourmaline pseudomorphs in the form of fine, silky, hair-like needles, intricately matted, or, in places, in radiating bunches.

Similar nodules have been described by Barrell [10] and Knopf [11] in aplite in Montana, and consist of quartz, tourmaline and orthoclase. Both authors consider their origin due to segregations in the liquid magma, although Barrell believes the presence of boron and fluorine in the tourmaline may indicate pneumatolytic action, whereas Knopf regards the tourmaline as purely pyrogenic. Other similar quartz-tourmaline nodules from Tasmania are described by Waller [12] and Waterhouse,[13] who explain them as due to magmatic segregation. Tilley,[14] however, believes they are of miarolitic origin, and refers to quartz-tourmaline nodules of undoubted pneumatolytic origin in aplite from Cape Willoughby, South Australia.

[10] Barrell, J.: Microscopical Petrography of the Elkhorn Mining District, Montana. U. S. Geol. Survey, 22d Ann. Rept., pp. 542–543, 1901.

[11] Knopf, A.: U. S. Geol. Survey, Bull. 527, pp. 34, 35, 53, 1913.

[12] Waller, G. A.: Report on the Tin Ore Deposits of Mt. Heemskirk. Tasmania Dept. of Mines, Sept. 1902, p. 4.

[13] Waterhouse, L. L.: The Stanley River Tin Field. Tasmania Geol. Survey, Bull. 15, p. 28, 1914. Also; The South Heemskirk Tin Field. Tasmania Geol. Survey, Bull. 21, p. 70, 1916.

[14] Tilley, C. G.: The Occurrence and Origin of Certain Quartz-Tourmaline Nodules in the Granite of Cape Willoughby. Royal Soc. South Australia, Trans., vol. XLIII, p. 156, 1919.

A comparison of the nodules from Montana, Tasmania, and South Australia with those from Llallagua-Uncia, suggests that the Bolivian examples are undoubtedly of pneumatolytic origin.

It is supposed [15] that one result of metasomatic action of the solutions upon the country rocks was the removal of calcium and all alkalies, which combined with the excess of boron to form soluble meta- and pyroborates with the addition of aluminum and fluorine. The released silica formed secondary quartz.

Either the tourmalinization process was of long duration, or there were several periods closely following each other.

Although such characteristic fluorine minerals as fluorite, topaz, and lepidolite are lacking, other indications show that boron was associated with fluorine in the magmatic solutions. Rare apatite and abundant wavelite, containing acid water, point to the presence of fluorine, and it is possible that the tourmaline also contains both boron and fluorine.

Kaolinization.—Where tourmaline is scarce, its place is taken by kaolinite, chlorite and muscovite. Minute hexagonal flakes of kaolinite form an opaque, earthy aggregate, replacing large phenocrysts and small grains of feldspar.

The assumption of weathering as the cause of kaolinization meets with the objection of the possible influence of descending waters upon so large a mass of rock at such depth. It is more likely that the replacement of feldspar by kaolinite was due to alteration by hypogene solutions. Although it is sometimes assumed that kaolin is rarely formed by hot waters at considerable depth, it has been shown that it is in many instances a result of metasomatic action by hydrothermal water.[16]

Silicification.—During the process of kaolinization, some silica was liberated and redeposited. Although the amount was in general negligible, some parts of the rock are strongly silicified. Such silicification might suggest the addition of silica, but in

[15] McAlister, D. A.: *Op. cit.*

[16] Ransome, F. L.: The Association of Alunite with Gold in the Goldfield District, Nevada. ECON. GEOL., vol. 2, p. 667, 1907. Zalinski, E. R.: Tourquoise in the Burro Mountains, New Mexico. ECON. GEOL., vol. 2, p. 464, 1907. Cornu, F.: Tscherm. Miner. und Petr. Mitt., vol. 25, p. 234, 1906.

the somewhat similar transformations of granite into the Gligga Head greisen,[17] into trowlesworthite,[18] and especially into luxullianite,[19] the amount of silica liberated from the breaking down of orthoclase and biotite was quite large, and it is doubtful whether much silica was added.

Other Features.—Microscopic study of such altered rock yields little information as to its original nature; it suggests that it may have been a dacite-porphyry or rhyolite-porphyry, but it is impossible to name with certainty a rock in which all feldspars have been altered. No chemical analysis was attempted because of the impossibility of obtaining fresh material.

A prominent feature of the porphyry is the presence of rock fragments and finely comminuted volcanic material. In some sections, there are inclusions of small rounded fragments of the porphyry itself or fragments of tourmalinized sandstone. Another feature is the complete absence of glass, which could hardly have remained unchanged under the powerful action of mineralizers. Commonly, tuff and breccia are so intermingled with porphyry that it is impossible to separate the three phases in the field.

Positive evidence for the intrusive origin of the Llallagua-Uncia rock is found in its shape and occurrence, but the rock possesses some features common to effusive rocks. Its vesicular and locally pumiceous character, the quartz phenocrysts fractured, evidently, by an excessively viscid magma, the accompanying pyroclastics, the lack of uniformity in texture and, finally, the felsitic groundmass, are all features testifying to surface or near-surface conditions of solidification. The magma penetrated higher into the crust than is common with intrusive bodies, filled the volcanic vent, and congealed in the cooler zone immediately under the surface.

Although some erosion must have taken place since the erup-

[17] Scrivenor, J. B.: The Granite and Greisen of Cligga Head. Quar. Jour. Geol. Soc. (London), vol. 59, p. 142, 1903.

[18] Bonney, T. G.: Roy. Geol. Soc. of Cornwall, Trans., vol. 10, p. 182, 1887.

[19] Bonney, T. G.: On the Microscopic Structure of Luxullianite. Miner. Mag., vol. 1, p. 215, 1877.

tive period, there is evidence that this process was not extensive. Erosion has swept away the volcanic cone, but did not remove even the whole scoriaceous crust of the plug.

Dikes.

The porphyry intrusion was accompanied by formation of a series of vein-like bodies ("bancos") within the margin of the plug and extending out into the graywacke. Although resembling regular igneous dikes in shape and other respects, they are irregular in form and have variable strike and dip. The dike fissures were the earliest superficial contraction cracks and form a single system unrelated to the later vein fissures. The veins usually meet the dikes obliquely, or, in a few instances, parallel them. The best example of the latter is the Contacto vein, with a dike of "mottled rock" accompanying it along one of the walls. The dikes themselves seem somewhat puzzling, not so much as to mode of origin, but as to the source of the material of which they consist, especially of the numerous included sandstone fragments. The latter form the main constituent of the dikes, being cemented by a fine porous matrix of tourmaline and quartz which evidently represents devitrified and altered glass.

The dikes are especially numerous in a belt limited by the 411 and 446 levels in the vicinity of the sandstone mass which forms the wall rock of some large veins. This suggests that the proximity is not accidental since, in thin section, the sandstone is seen to be a clastic rock (Fig. 4) composed of quartz and tourmaline particles of different sizes, clearly erosional detritus from porphyritic material such as filled the plug. The above explanation coincides with other evidence relative to the origin of the sandstone, and its age as compared with the volcanic rock. On the 446-meter level, the contact between the porphyry and the sandstone is exposed; the former contains fragments of dark gray sandstone, and porphyry fragments are embedded in the sandstone.

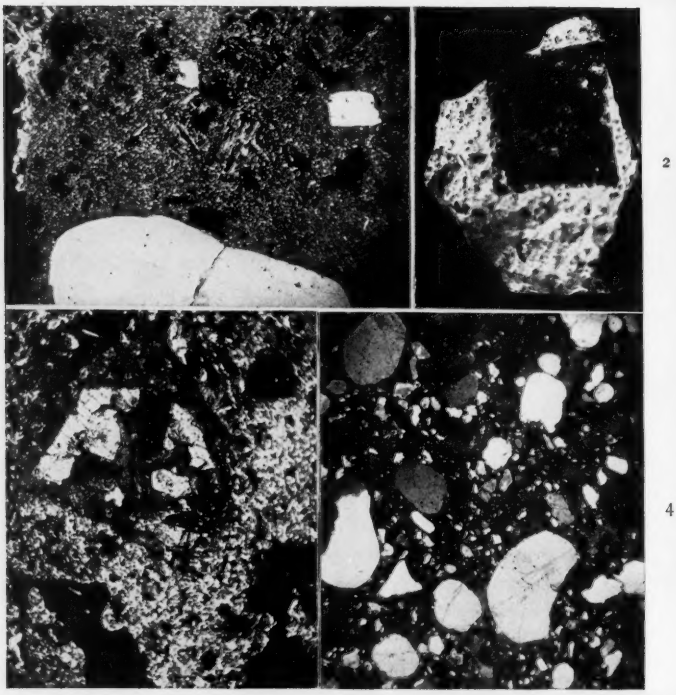

FIG. 1. Porous tourmalinized porphyry near Rosas vein, level 446, Uncia. Quartz phenocrysts partly resorbed, partly with preserved form; groundmass of tourmaline and quartz; veinlet of wavelite in upper left corner. X-nicols, × 27.

FIG. 2. Large tourmaline-quartz phenocryst pseudomorph with form of host crystal, in altered porphyry. Near Reggis vein, level 446, Uncia. × 1½.

FIG. 3. Large pseudomorph of granular tourmaline aggregate after feldspar, in altered porous porphyry near Transformadora vein, level 516, Uncia. X-nicols, × 38.

FIG. 4. Tourmalinized and silicified sandstone near Branch 67, Forastera vein, level 446, Uncia. Coarse, rounded quartz in a fine-grained matrix of comminuted tourmaline and quartz. X-nicols, × 27.

THE VEINS.

The most striking feature of the fissures is that they are almost all confined to the area of the volcanic plug; only a few important fractures occur in the graywacke and sandstone. The fissures belong to three distinct systems; one is represented by small veins; another by groups of larger veins that occupy fault fissures; and the third by non-mineralized faults.

In the upper part of the volcanic neck there are numerous small veins that vary from one to fifty centimeters in width. At depth they die out, or join the major veins. They are narrow, lack evidence of movement, and are probably contraction cracks due to cooling and settling of the volcanic mass. These veins are seldom persistent in strike or dip although they tend to parallel the larger veins. The mineral content varies in different veins, and even in the same vein; some contain only barren sulphides. In general, these smaller veins are not as rich as the main veins but, due to their great number, they are of considerable economic importance. The walls show some replacement.

The mineralization of these minor fissures probably did not occur until after they had been cut by a younger system of fault fissures, less numerous but more persistent, which served as feeders of metal-bearing solutions, both for this group and for the smaller branch veins. The most important, such as the San Fermin-San Jose, Forastera, Contacto, Reggis-Inca, and Salvadora veins, belong to this second system of fault veins. Their width ranges from a meter or less up to 6 or 7 meters. Brecciation and relatively great displacement of the walls characterize them. Few reach the surface, but they are persistent in strike and dip: The strike varies from N.15E. to N.70E.[20] and the average dip is about 70°. The fillings are sharply defined from the country rock.

The mineralization of the two productive fissure systems was followed by strong movement, resulting in many faults. All the fissures of this third group are barren. Only one series, which

[20] Koeberlin, F. R.: Geologic Criteria for Large-Scale Tin Prospecting in Bolivia. Eng. and Min. Jour., vol. 123, p. 280, 1927.

includes the large Stanton and Diaz faults, striking N.10W., have cut and displaced large veins of the second epoch.

Paragenesis of Minerals.

Tourmaline was deposited first, replacing other minerals, filling cavities, and forming inclusions. It is abundant in the country rock up to the top of the volcanic plug, increasing noticeably downward, but it is scarce in the vein fillings, where it occurs only as occasional inclusions in quartz. The common mode of occurrence is as minute grains, of very pale colors, without pleochroism and lacking crystal form. All this causes difficulty in the determination of this peculiar variety of tourmaline.

Under the microscope in ordinary light, the colors vary from light brown to brownish-gray, rarely to brownish-green or bluish-green. The mineral appears to be optically negative ($2V = -0$). The index of refraction is $\omega = 1.63$. The double refraction $\omega - \varepsilon = 0.023$. The pleochroism is generally feeble, though distinct in some grains. In several thin sections the tourmaline may be confused with sericite, but the negative elongation, higher index of refraction, stronger absorption of light perpendicular to the principal zone, and variability of color, are diagnostic.

From its mode of occurrence in the porphyry, it has the aspect of a secondary mineral closely associated with the decomposition of feldspar and biotite. The composition of tourmaline is complex and variable, and would suggest that under certain circumstances it might occur as the decomposition product of more than one mineral.[21] In the porphyry, the tourmaline evidently does result from post-solidification effects due to the concentrated action of the most volatile constituents of the magma on feldspar and biotite.

The inclusion of microscopic tourmaline in quartz, in places oriented parallel to the prism and pyramid planes of the crystal, as described elsewhere, is common. But in view of the positive evidence of the primary origin of such quartz crystals as original

[21] Bonney, T. G.: Miner. Mag., vol. 1, *loc. cit.*

phenocrysts, and on the hypothesis that tourmaline is a secondary mineral, this occurrence seems perplexing. In attempting to explain a similar occurrence in dikes in Tasmania, Weston-Dunn [22] suggests that they were still molten and that quartz phenocrysts continued their growth during tourmalinization. Such an explanation would, of course, be untenable in this case. Another hypothesis is put forward by Bonney,[23] who believes that the tourmaline-yielding agents rendered quartz sufficiently plastic to yield to the force of crystallization. Hydrofluoric acid will gelatinize quartz. Although fluorine was present in the emanations, the amount was small and hardly enough to produce softening of the quartz. According to Lindgren,[24] "it is not necessary to assume complete permeation and softening of a rock . . . in order to account for secondary developed crystals. The phenomenon may be caused by simple metasomatic replacement."

It has been established that tourmaline is always a deep-seated high-temperature mineral, and its association with tin ores indicates plutonic conditions. But the recognition of this mineral as deep-seated seems contradicted by the field relations of the volcanic plug. Brammall and Harwood [25] have attempted to determine the temperature-range of the formation of tourmaline. In geodes in granite, tourmaline is demonstrably contemporary with α-phase quartz and evidently formed below 575° C., but how far this temperature-range may extend downward is uncertain. Grout [26] gives the formation temperature for small crystals of tourmaline as below 300° C. Extremely minute, pale-greenish tourmaline prisms of this type have been observed in Skiddaw granite and in certain Bolivian phyllites.

Quartz of secondary origin resulted from the replacement of some of the porphyry minerals by tourmaline with the simultane-

22 Weston-Dunn, J. C.: The Economic Geology of the Mount Bishoff Tin Deposits, Tasmania. Econ. Geol., vol. 17, p. 153, 1922.

23 Miner. Mag., vol. 1, *loc. cit.*

24 Lindgren, W.: Metasomatic Processes in Fissure Veins. Amer. Inst. Min. Eng., Trans., vol. 30, p. 597, 1901.

25 Brammall, A., and Harwood, H. F.: The Temperature Range of Formation for Tourmaline in the Dartmoor Granite. Miner. Mag., vol. 21, p. 205, 1926.

26 Grout, F. F.: Petrography and Petrology, p. 441. New York, 1932.

ous release of large quantities of silica. Redeposition together with tourmaline occurred partly in the space formerly occupied by the replaced minerals, and partly in the groundmass. The first part of this secondary fine-grained quartz is so intimately associated with the tourmaline that there can be no doubt as to its source. It was clearly released during the destructive replacement of feldspar and biotite.

Vein quartz of α-phase is the commonest vein filling, with the exception of the sulphides. It is commonly coarse-banded, crustified, and variable in size and habit of crystals, occurring generally in druses, as long, colorless, idiomorphic crystals against the later minerals. Its deposition began near the close of the tourmalinization period, as indicated by rare inclusions of tourmaline needles in quartz crystals, which also afford positive proof of the overlapping deposition of both these minerals.

Cassiterite occurs in compact masses of closely interlocking grains, such massive texture being a characteristic feature of Bolivian cassiterite.[27] Another common occurrence is as a filling in the interstices between quartz crystals. More rarely, cassiterite occurs as small, short, pyramidal crystals in druses. In addition, it has been introduced into the country rock in small quantities.

Under the microscope, the cassiterite is of comparatively light color in some places, ranging from brownish-yellow to brownish-gray, but the very dark, almost black, variety is most typical of the Llallagua-Uncia deposit. Another characteristic is a distribution of color in zones paralleling the pyramidal faces. Locally, the cleavage parallel to (100) is noticeable and this and the zonal coloring are the most important means of distinguishing twins and separate grains in aggregates. In polarized light, "knee-twins" are seen to be very widespread and to follow the Zinnwald law, with the twinning plane parallel to (101). The high index of refraction, strong double refraction, and high interference colors are distinctive.

The cassiterite belongs to the pneumatolytic type, according to

[27] Beck, R.: Erzlagerstätten, 3d ed., vol. I, p. 222. Berlin, 1901.

Ahlfeld.[28] An analysis by Greene[29] gave 94.06 per cent SnO_2 and 4.30 per cent Fe, and transparent crystals of "pure" cassiterite from "high-temperature" veins of Araca gave 100 per cent SnO_2. The cassiterite with high iron content is believed to form under conditions of comparatively low temperature.[30]

Banding in the veins may indicate several generations of quartz, cassiterite and iron sulphides; however, it is reasonably certain that in general this quartz began to form after the tourmaline, and before the cassiterite.

Arsenopyrite, found only in the lower levels, is undoubtedly one of the earliest sulphides. It may conceivably be of the same age as pyrrhotite, but no definite relations could be established.

Wolframite probably formed under varying conditions, as indicated by its range of composition. *Scheelite* in colorless crystals occurs associated with wolframite.

Bismuthinite is abundant, especially in the northern part of the deposit, as a regular associate of cassiterite, in crystals several centimeters long, or more rarely in fine-grained aggregates. The mineral is younger than the cassiterite, although Davy[31] believes that both are nearly of the same age as the native bismuth.

Native Bismuth is less abundant than bismuthinite, although in the upper levels it commonly occurs inclosed in bismuthinite. Kittle[32] believes that it is a product of reduction of bismuthinite.

Iron Sulphides are mainly developed from pyrrhotite, although some pyrite is evidently hypogene and is undoubtedly of younger formation than cassiterite. The alteration of pyrrhotite to pyrite and marcasite is probably metasomatic, as pseudomorphs of pyrite

[28] Ahlfeld, F.: Die Erzlagerstätten in der tertiären Magmaprovinz, etc. Neues Jahrb. für Min. Beil. Bd. 65, Abt. A, p. 333, 1932.

[29] Greene, G. U.: Solubility of Tin Minerals. Eng. and Min. Jour., vol. 122, p 417, 1916.

[30] Ahlfeld, F.: Contribucion al conocimiento de las contracciones secundarias de metales con los yacimientos bolivianos. Revista Minera de Bolivia, vol. 5, p. 147, Mayo, 1927.

[31] Davy, M.: Ore Deposition in the Bolivian Tin-Silver Deposits. ECON. GEOL., vol. 15, p. 493, 1920.

[32] Kittle, E.: Los Yacimientos Estanniferos de Bolivia. Revista Minera de Bolivia, Marzo, 1928.

after pyrrhotite are found. Apparently the latter was attacked by alkaline polysulphide solutions. The preferential alteration to pyrite or marcasite is regarded as dependent on temperature.

The iron sulphides are the commonest of the vein minerals. Generally the pyrite and marcasite are massive, granular, botryoidal or stalactitic but they occur commonly as well developed crystals in druses.

Pyrrhotite was deposited later than the cassiterite, although its formation may have partially coincided with that of cassiterite.

Stannite is the most characteristic ore mineral, and probably the first deposited during the hydrothermal phase. It is common in certain veins in the upper levels, but is never seen in depth. The usual occurrence is massive or in granular masses, but in some veins it occurs in druses as tetragonal twinned crystals. Only a small percentage of the specimens have the typical steel-gray color of the Cornish stannite; the major part are brown, resembling zincblende, and give a brown streak, but are without metallic luster. It fuses with difficulty, sometimes barely rounding the edges, and is very porous and brittle. All these unusual features are characteristic of the stannite from the Llallagua-Uncia deposit. According to Ahlfeld,[33] brown stannite from Vila Apacheta contains zinc, the zinc possibly replacing a part of the iron.

Sphalerite occurs as the black or reddish-black variety, rich in iron, forming bands or large scattered crystals. It deposited during the hydrothermal phase, but its exact paragenetic position is unknown.

Supergene Minerals. The oxidation of iron sulphides yielded alteration to " limonite " in the form of a yellow ocher or a dark, slaggy-looking crust. The zones of deposition of stannite are always marked by secondary copper minerals resulting from the decomposition of this mineral. Chalcocite, covellite, bornite, native copper and, above all, an abundance of copper sulphate in the form of stalactites, give positive evidence of the presence of stannite. Tin released from oxidation of this sulphostannate

[33] Ahlfeld, F.: *Op. cit.*, p. 338.

mineral possibly was deposited partly as supergene cassiterite. Although stannite is common in this deposit, it could hardly be admitted that a large amount of supergene cassiterite would result from this weathering. Available evidence of redeposition of cassiterite as supergene in appreciable quantities, and inadequate criteria for distinguishing supergene cassiterite from hypogene, do not make a conclusive case.

Phosphates attract unusual scientific interest in the Llallagua-Uncia deposit, since they are present in great quantity as in no other tin deposit. Several supergene phosphates of iron and aluminum occur locally in abundance and indicate that phosphorus was given off in considerable quantities by the magma. Of the hypogene phosphates, apart from rare occurrences of monazite, only microscopic apatite intergrown with cassiterite was observed. According to Spencer,[34] the comparative rarity of apatite may be accounted for by its inconspicuous and unusual mode of occurrence and its ready solubility in acid water. The original apatite content of the ore was probably much higher, but decreased as the supergene phosphates formed at the expense of the primary mineral. Of the secondary phosphates, the most abundant is wavelite, which commonly occurs as sphero-crystals or as incrustations on cassiterite, quartz and marcasite. The strongly acid reaction in the test for water shows that the mineral contains fluorine. Next in abundance comes vivianite, a highly characteristic mineral of the Bolivian tin deposits. It occurs in druses of well developed crystals, commonly twinned, or filling veinlets. Finally, there are the unique minerals described by Gordon,[35] vauxite, metavauxite, and paravauxite, none of which have been observed elsewhere.

ORE GENESIS.

Several tin deposits in the southern part of the Bolivian tin belt are restricted to comparatively small, separate, but probably related, intrusions of porphyry. Not only do the Llallagua-Uncia veins appear to be genetically related to the rock inclosing them

[34] Spencer, L. J.: *Loc. cit.*

[35] Gordon, S.: Acad. of Nat. Sci. Phila., Proc., vol. 76, p. 325, 1924.

through their common source, but the peculiar character of mineralization is undoubtedly closely related to the nature of the rock and its conditions of emplacement.

Although some of the minerals suggest high-temperature formation, other evidences point to deposition at lower temperatures near the surface. The temperatures of deposition are dependent not merely upon the heat furnished to the ascending solutions by the inclosing rock, but also upon the heat supplied by the magmatic source furnishing the solutions, and upon other factors. These conditions explain the apparent anomaly in the mineralization of the Llallagua-Uncia deposit.

There is evidence that after cooling and fissuring of the volcanic neck, emanations, probably gaseous, rich in water, in some metals, and especially in boron and phosphorus, ascended to regions of lower pressure and encountered areas of highly porous, cracked, and shattered rock. A part of the hot gases pressed rapidly outward, easily penetrating beyond the actual walls of the fissures, producing notable changes in the rock, and finally escaped. With the release of a large portion of the volatile components, there would be an inevitable reduction in pressure. For many components, the new pressure would be below the critical point, and under conditions existing in the cool upper horizons of the fissures, they would revert to the liquid state.

From a comparison of the mineralization in the altered country rocks and in the veins, it is certain that the extensive metasomatic alteration of the rocks was largely antecedent to, rather than contemporaneous with, vein formation. The alteration of huge volumes of rock is of a character indicating permeation by a gaseous agency, whereas the filling of the fissures, with restricted changes in the wall rock, suggests deposition from liquid solutions. These facts lead to the conclusion that the concentration of some mineral components caused deposition of minerals in the vein fissures, and the condensation of gaseous solutions was the result of dissemination and loss of the bulk of the volatile components through the porous mass.

If the above hypothesis of the escape of the volatiles is sub-

stantially correct, the important chemical and physical factors responsible for the differences in the material deposited in the country rock and in the different zones of the veins can be explained. Owing to the loss of part of the volatile constituents—including the most active mineralizers—there would be a decrease in acidity of the original magmatic extracts, with a progressive concentration of the other components.

The common effect of magmatic solutions is the formation of minerals rich in the most concentrated elements of those solutions. Among those of the latter which formed the Llallagua-Uncia veins, boron was especially abundant. This element readily forms complex volatile compounds that may give rise to the formation of new combinations [36] and possibly take some part in extraction of tin dioxide from the magma; [37] so naturally the first effect was deposition of the boron mineral tourmaline, and cassiterite. The initial temperature of the solutions must have been fairly high to have allowed tourmaline to form, but near the surface, temperature and pressure must have decreased. The pneumatolytic phase still persisted at a somewhat lower temperature, but the new conditions were apparently favorable to the development of other minerals rather than tourmaline.

Possibly at this stage, the deposition of quartz, cassiterite and some sulphides began in the fissures, while the deposition of tourmaline was rapidly decreasing. The former minerals belong to a distinctly later period than tourmaline, since there are only a few inclusions of the latter in vein quartz. Cassiterite is limited to a definite zone. The lower limits are encountered in the lowest levels of the mine (around 700 m.), and the upper boundary extends in some places close to the top of the volcanic neck. Because of the high temperature, the solutions could easily reach the upper zones of the fissures before reverting to the liquid state. There, owing to the loss of the most active ingredients and changes in physical conditions, the composition and state of the solutions underwent rapid changes. When they were close to the surface, the whole of their metallic load was thrown down, not in a

[36] Vogt, J. H. L.: On Ore Magma. Eng. and Min. Jour., vol. 123, p. 648, 1927.
[37] McAlister, D. A.: *Op. cit.*

normally developed zonal arrangement, but in rapid succession with much overlapping and superimposition. Support for this conclusion is found in the upper zone of the deposit, where complex ores consisting of small amounts of cassiterite with sulphides and sulpho-salts of tin, lead, iron, zinc, copper and antimony, are found within a few hundred meters of the surface. These ores are so intimately intermingled that it is reasonable to infer a nearly simultaneous deposition. Perhaps stannite commenced to deposit first, but it is not possible to give a definite succession of minerals during this hydrothermal phase. It is also difficult to establish the level to which ore deposition extended upward, because oxidation and leaching have removed the upper portions of the veins, but it is probable that the actual deposition reached higher zones than is indicated by copper and zinc sulphides. Stelzner [38] mentions the presence of an abundance of silver ore in the Blanca vein, and the recently opened Callaperia vein affords another indication that the lead-silver zone is feebly represented.

CONCLUSIONS.

Since Stelzner first described the Bolivian tin deposits, several attempts have been made to compare them with the normal tin veins of Cornwall and Erzgebirge. Stelzner himself considered the Bolivian veins different mineralogically and genetically from those of other countries, and therefore constituting a separate type which he called "Typus Potosi." Later, Steinmann [39] drew attention to another part of the country where tin veins occur with a different mineralization and an absence of silver, and these, in his opinion, may be regarded as another type which he named "Typus Araca."

The veins in Bolivia show different mineralogic composition such as would represent two distinct vein types, but since the genetic relation of all these deposits through a common source has been established by Steinmann, Singewald, Armas, and others, they must be considered together as members of a continuous

[38] *Op. cit.*, p. 129.

[39] Steinmann, G.: Über die Zinnerzlagerstätten Bolivias. Monatberichte der Deut. geol. Ges., p. 7, 1907.

series rather than representing two different types. In other words, the high-temperature deposits grade into those deposited under more moderate conditions, with a corresponding change of predominant minerals. Some deposits clearly show this transitional character. Of these the Llallagua-Uncia deposit constitutes the best example. In one part of this deposit, ores are observed with the characteristics of ores from northern Bolivia; that is, ores formed under high-temperature conditions, with normal tourmaline-cassiterite mineralization of a plutonic character. In another part, the veins are of hydrothermal origin, with associated minerals characteristic of the cooler zone, such as are found in the southern section of the Bolivian tin belt.

In addition to the peculiar nature of its origin, the Llallagua-Uncia deposit possesses many features not observed in any other tin fields. The presence of phosphates in considerable quantity, the occurrence of a rare germanium mineral, and finally, an extraordinary quantity of tourmaline containing boron almost to the exclusion of fluorine, make this deposit unique among tin fields.

NEW YORK, N. Y.

DISCUSSION AND COMMUNICATIONS

ADIRONDACK MAGNETITE DEPOSITS.

Sir: In 1928 this Journal published my paper "Certain Magmatic Titaniferous Iron Ores and Their Origin," giving the results of study of deposits in the Adirondacks, Quebec, and Ontario. I concluded that some of the concentrations could be best explained as filter-pressed, orthotectic differentiates of the associated parent gabbro or anorthosite. Since publishing this paper, I have made additional observations on the titaniferous ores and anorthosite in the Morin and Saguenay districts of Quebec and have realized some difficulties in the application of my theory if, as contended by some Adirondack geologists, syenite is also a restmagma from the crystallization of the parent magma of the gabbro and anorthosite.

Bowen, Buddington, and Balk have emphasized the probability that the anorthosite of the Adirondacks is the result of the accumulation of crystals leaving a syenitic restmagma, which now forms independent massifs. It would be improbable if not impossible to have together a syenitic restmagma and one enriched in the iron ore minerals. Cushing and Miller believe that the Adirondack syenite is a separate intrusive series, and I believe that this is the case in the Morin and Saguenay districts. There is no trace of syenitic interstices in the anorthosite or even in the chilled border facies, which is coarse-grained and ophitic where it is exposed north of Morin Heights in the Morin district. South of Lac St. Jean I examined a part of the roof of the Saguenay anorthosite where it has a low dip and found syenitic material lacking in the gabbroid anorthosite, which is the roof facies.

A very few pegmatite dikes containing some quartz cut the anorthosite, and are later than the more numerous basic dikes.

In the Morin area no syenites are known to be related to the anorthosite, but pyroxene granites, syenites, and diorites, which Adams [1] called the pyroxene granulites and showed to be older than the anorthosite, are common. I believe that the part of the Buckingham series consisting of pyroxene granites and syenites described by Wilson [2] is equivalent to the pyroxene granulites and older than the anorthosite of the Morin and Saguenay areas.

The widely-developed Roberval series named and described by Dresser [3] is younger than the anorthosite, and contains syenitic members but is probably unrelated to the anorthosite. If the syenite of the Adirondack area is also unrelated to the anorthosite, one of the difficulties in the way of the acceptance of my theory for the titaniferous ores is removed.

Some deposits of iron ores associated with gabbro and anorthosite are certainly of pneumatolytic or hydrothermal origin. It is interesting to speculate on a possible source of the material for the iron-ore minerals by concentration by solutions of the schiller inclusions in the plagioclase. As is well known, ilmenite and magnetite are found along with spinel and pyroxene in the very minute but numerous inclusions within the dark-colored feldspar of the anorthosite. As a rule, where the plagioclase has been sheared, the inclusions have been collected into larger grains or removed in solution. Although the amount of such inclusions in one crystal is small, nevertheless shearing with the decolorization of the feldspar is widespread in these rocks, and considerable material might be added to the solutions to form magnetite and titanite in a metasomatic deposit. This suggestion is not offered as an alternative explanation for the deposits mentioned in the first part of the note, but for some replacement deposits related to the gabbroid rocks.

F. Fitz Osborne.

McGill University,
Montreal, P. Q.

1 Summary Report of the Geological Survey of Canada for 1895.
2 Geological Survey of Canada, Mem. 136.
3 Geological Survey of Canada, Mem. 92.

TALC AS A LUBRICATOR FOR POLISHED SURFACES MOUNTED IN BAKELITE.

Sir:—An annoying and time-consuming nuisance encountered in mounting specimens in bakelite is the tendency of the bakelite to stick to the steel parts of the mounting device. Krieger and Bird[1] recommended that these parts be highly polished to facilitate removal of the specimen. Fuller[2] has described a refinement of the molding apparatus which simplified the removal of the base, specimen, and plunger from the sleeve. There remains some danger of harming the mount when it is disengaged from the base and the plunger. In spite of polishing, continued use of these parts results in a tendency to stick to the bakelite.

This difficulty can be overcome by applying a thin film of pure talc to the parts before each operation. The writer applied the film by first rubbing the surface with a lump of talc and then wiping off the excess talc with the finger. The talc does not detract from the appearance of the mount, since little of it is required. There is no effect except the desired one of lubrication. When talc is used, the base and plunger need not be polished, a good machining job being sufficient. It was further found that a dusting of talc on the inside of the sleeve seems to lessen the adherence of the bakelite to that part.

PHILIP B. MYERS.

LEHIGH UNIVERSITY,
BETHLEHEM, PA.

[1] Krieger, Philip, and Bird, Paul H.: Mounting Polished Surfaces in Bakelite ECON. GEOL., vol. 27, pp. 675–678, 1932.

[2] Fuller, Harry C.: Mounting Polished Surfaces in Bakelite. ECON. GEOL., vol. 28, pp. 393–395, 1933.

REVIEWS

Dip and Strike Problems. By K. W. EARLE. Pp. x + 126, figs. 122, pl. 1. Thos. Murby and Co., London, 1934. Price, 12 s. 6 d.

The author furnishes trigonometrical formulas for the solution of problems involving the use of dip and strike data, both accurately and rapidly, by means of the slide rule. He prefers for this purpose formulas rather than geometrical constructions, because different constructions are necessary for each problem and the solutions arrived at by their use are only approximate.

Each formula proposed is derived from assumed data, such as are obtained by observations of dips and strikes at various points in a simple or a folded series of beds, on fault traces, etc., and its application is illustrated by the solution of a problem. The most interesting solutions are those relating to the dips and strikes of underground beds and veins, to the srtuctural elements of folds, to the throw and heave of faults in parallel beds, from surface and from underground observations, and in anticlines and synclines, from the displacements of surface outcrops on opposite sides of the fault trace. The book closes with a chapter on a few mechanical devices for solving simple dip and strike problems rapidly.

W. S. BAYLEY.

Physical Geology. By C. R. LONGWELL, A. KNOPF, and R. F. FLINT. Pp. 514, figs. 341. John Wiley and Sons, Inc., New York, 1932. Price, $3.75.

Outlines of Physical Geology. By C. R. LONGWELL, A. KNOPF, and R. F. FLINT. Pp. 356, figs. 297. John Wiley and Sons, Inc., New York, 1934. Price, $3.00.

"Outlines of Physical Geology" is a brief edition of the larger book, and carries the same conspicuous values as well as the same occasional shortcomings. Especially commendable in both textbooks are the generally new and effective illustrations. Only a few, such as Figure 22 in the "Outlines," are not wholly convincing. The captions are complete and thus make the figure an integral part of the text. As textural presentation, the headings and subheadings are also noteworthy: they form a useful, logical outline of the matter presented.

Certain minor shortcomings may be recognized, although probably attributable to the fact that such books are designed to serve beginners

with interests so varied that they range from the philosophical to the purely technical aspects of the science. Thus, adequate footnotes are lacking,—a concession perhaps to the average freshman. Moreover, a fundamental concept such as uniformitarianism might have been emphasized for its historical significance and mentioned by name as well; where, as on pages 2 and 8 of the "Outlines," it is touched upon, the doctrine might have been connected with the name of its great enunciator.

In substance, the concreteness with which fundamental processes are handled is exceptional. The "case method" is used with conspicuous success. As illustration, the discussion of weathering at Durham Castle ("Outlines," pp. 12–13) may serve.

The style, too, is lucid, in places almost epigrammatic. Thus, "The portion of the Earth's history that is written in the rocks is referred to as geologic time." That statement is clear and to the point, and at the same time it gives the student pause for reflection.

The appendix on topographic maps is a welcome and useful addition. So also are appendices on minerals and rocks, though there is question whether the text might not have been boldly begun with chapters on minerals and rocks,—the earth materials. Such a procedure would obviate the problem of discussing rock materials in a preliminary fashion first and again later in detail. For textbooks to be used in American universities, this approach is a moot question, partly because in systematic geology we cannot bring ourselves to do otherwise in the teaching of geology than to "start from scratch," neglecting the necessary chemical and mineralogical background. Ultimately we shall have to frankly recognize two kinds of beginners' textbooks. One kind will assume chemical and mineralogical background. The other kind will build up the necessary knowledge of these more fundamental disciplines as the authors progress. Of the latter type the two texts under review are admirable and highly commendable examples. In finish and care they leave nothing to be desired in any important respects.

CHAS. H. BEHRE, JR.

NORTHWESTERN UNIVERSITY,
EVANSTON, ILL.

Les Resources Minérales de la France d'Outre-Mer. I. Le Charbon. By F. BLONDEL. Pp. 245, figs. 33. Soc. d'Éditions Géographiques, Maritimes et Coloniales. Paris, 1933. Price, $2.25.

A few months ago attention was called (this Journal, vol. 28, p. 186) to the series of volumes designed to bring to the notice of the public the resources of the French Colonies. The first volume gave a general view of the geology and mining resources of these colonies. This was to be followed by other volumes dealing in detail with special products. The

first of these, as mentioned above, deals with the occurrence of coal in all French colonies and its relation to the economic welfare of France. The three centers most important in the French Economy are in Indo-China, which produces annually about 2,000,000 tons of anthracite, the Kénadza district in Algeria, which furnishes about 20,000 tons of a volatile product, and the Djérada district in Morocco, which yields about an equal tonnage of anthracite. The other sources within the French group of colonies are unimportant. They consist mainly of comparatively small deposits of lignite.

In the present volume, L. Clariond gives a brief summary of the geology of the coal areas in Tunisia, Algeria, and Morocco, illustrating his discussion with sketch maps of the geology of the various producing districts, and cross-sections of the various occurrences. From a study of the geology of these districts, it appears probable that in the vicinity of Djérada and Kénadza there are several tens of millions tons of coal that will be available for more rapid exploitation than is now possible, when means of transportation are better developed.

The description of Madagascar occurrences is undertaken by F. Blondel, who sketches the geology of the island, with special reference to its stratigraphy. The principal occurrences of coal now available are in a series of basins in the southwest part of the island. In the Sakoa basin alone it is estimated that there are workable at least 350,000,000 tons of a fair quality of steaming coal.

The bulk of the volume, which is by Blondel, is an account of the coal resources of Indo-China. Here there are a great many deposits in two groups, one of which consists of beds of anthracite of upper Triassic, or Rhaetic, age and the other of lignite beds of Tertiary age, occurring in many small isolated basins. The principal anthracite deposits are in the vicinity of the mouth of the Tonquin River, the most important being near Quang-yen. Recently a carboniferous deposit has been reported from Saravane, about 700 kilometers farther south. Summaries of the geology of each of the districts are given, illustrated by sketch maps and cross sections of the productive areas, which show clearly that the coal is in folded rocks and that the geology of the areas is complicated.

The principal lignite occurrences are in the neighborhood of Tuyen-quang on a branch of the Red River. Here the annual production is about 26,000 tons. Other occurrences are at present of little importance, although several of them yield annually from 1000 to 4000 tons.

The total production of coal in Indo-China during 1932 was 1,714,000 tons, of which nine-tenths was anthracite. Of this total, 1,147,000 tons was exported, mainly to China and Japan.

A short resumé of the geology and coal resources of New Caledonia,

by L. Guilemot, completes the volume, which is supplied with an index of geographic names and another of geological and mining terms.

W. S. BAYLEY.

Les Ressources Minérales de la France d'Outre-Mer: II, Le Fer, le Manganèse, le Chrome, le Nickel, l'Étain, le Tungstène, le Graphite, le Glucinum, le Molybdène, le Cobalt, le Titane, le Vanadium. By G. BÉTIER, F. BLONDEL, L. NELTNER AND OTHERS. Pp. 436, figs. 59 Soc. d'Editions Géographiques, Maritimes et Coloniales. Paris, 1934. Price, $3.00.

This is the second volume dealing in some detail with the mineral resources of the French colonies.

The present volume, which is of the same general character as that on coal, is an excellent summary of the production and reserves of the minerals designated in its title, *i.e.* those yielding constituents employed in steel making. The greater part of the book is devoted to the discussions of iron, nickel, manganese and tin, but large parts also deal with chromium, cobalt and graphite.

The ores of iron are mainly in Morocco, Algeria and Tunisia, the reserves in these countries being respectively 60,000,000, 160,000,000, and 30,000,000 tons. In addition, the reserves in Spanish Morocco are estimated at 50,000,000 tons. Most of the material, and nearly all that being worked, is hematite occurring in veins in which are present also small quantities of sulphides. The ore is believed to be derived from pyrite and siderite. In some deposits magnetite is associated with the hematite in notable quantity. The discovery by geophysical methods of other deposits in Northern Africa is thought to be probable. The geology of the deposits, their annual yield of ore in recent years, and their economic importance are discussed in satisfactory detail. Most of the ores are of Bessemer grade. Their entrance into the world's trade is handicapped by high freight charges. Laterites, hematite, and magnetite are known to occur also at other points in French Africa, Madagascar, Indochina and other colonies, and at some places they have been exploited on a small scale.

Manganese ores are known in the Gold Coast and in the Soudan, also in the Cambrian and Permo-Triassic in Morocco. The deposits in Morocco are described in detail, their discussion being illustrated by sketch maps and sections showing their geological relations.

The most important chromite deposits are in New Caledonia, but others are known also in Indochina and Togo, and possibly in commercial quantities in Madagascar, Morocco and Algeria.

The occurrences of nickel ore in New Caledonia and at Sudbury are described in outline, but the most interesting feature of the chapter on

nickel is the excellent summary it gives of the properties and uses of the metal.

Another interesting chapter is that on tin ore, in which is given a brief, but satisfactory resumé of the methods of occurrence and origin of tin and tungsten minerals. The deposits in Indochina are discussed in a special chapter, and for comparison are outlined the important occurrences in Yunnan, China, and in briefer form all other productive occurrences elsewhere in the world.

Only a few pages deal with amorphous graphite, since only a few unimportant occurrences are in districts allied with France. Under the crystalline variety, the deposits in Ceylon and Madagascar are described more particularly with reference to the preparation of the material and its uses.

The other minerals named in the title of the report are dealt with lightly. Some are of relatively little importance and others, though important, are not found in commercial quantities in the French colonies. Beryl is worked on a small scale in Madagascar, and molybdenite in Morocco. The latter mineral occurs also in Madagascar, Indochina and New Caledonia, but only in small quantity so far as is now known. Vanadium minerals are known in small quantities in French possessions only in North and East Africa.

The occurrence of cobalt ores (asbolite) associated with nickel and chromium ores in alteration products of serpentine in New Caledonia, of smaltite veins near Marrakesh in Morocco and of linnaeite and heterogenite associated with the copper ores in Katanga, in addition to the well known deposits of Sudbury, suggests an abundant supply of sources of the metal, and the deposits of titaniferous sands in Senegal, in Reunion, Southern Annam, and Guadalupe are believed to offer abundant additional sources of ores of titanium.

W. S. BAYLEY.

Metamorphism: A Study of the Transformations of Rock Masses. By ALFRED HARKER. Pp. ix + 360, figs. 185. Methuen & Co., London, 1932. E. P. Dutton & Co., New York, 1934. Price, $5.90.

The distinguished author treats in this book of metamorphism, not of metamorphic rocks. He conceives of it "not as a status, but as a process." Consequently, detailed petrographic descriptions have been reduced to a minimum. Nevertheless, there are 185 double or triple drawings of rock sections as viewed microscopically. Part I deals with Thermal Metamorphism, and its nine chapters embrace general considerations, structures of rocks, thermal metamorphism of calcareous and non-calcareous sediments and of igneous rocks, and pneumatolysis and metasomatism. Part II (ten chapters) deals with Dynamic and Regional

Metamorphism, and treats of stress; dynamic and regional metamorphism; structures of crystalline schists; regional metamorphism of calcareous, non-calcareous, and igneous rocks; metamorphism with plutonic intrusion; repeated metamorphism; and retrograde metamorphism.

The individual subjects listed are covered adequately but unfortunately no mention is made of the newer investigations of Sander. The book would be of wider interest if the examples had not been chosen mainly from British sources. It is, however, a book that should be read by everyone interested in metamorphism.

BOOKS RECEIVED.

DAVID GALLAGHER.

Economic Geology of the Fife Coalfields. Area II, Cowdenbeath and Central Fife. J. K. ALLAN AND J. KNOX. Figs. 14; pls. 2. H. M. Stationery Office, Edinburgh, 1934. Physical features and summary of the geology. Descriptions of the Calciferous Sandstone, Upper and Lower Limestone Groups, Millstone Grit, and the coal-bearing strata; also a glossary of mining terms.

Moose Pass-Hope District, Kenai Peninsula, Alaska. R. TUCK. Pp. 61, pl. 1, figs. 6. U. S. Geol. Surv. Bull. 849–I, 1933, 15c. Regional geology; history, organization, geology, and production of gold lodes and placers.

Mineral Industry of Alaska in 1932. P. S. SMITH. Pp. 91, figs. 3. U. S. Geol. Surv. Bull. 857–A, 1934, 10c. Production and other economic data.

Geology of the Part of the Attock District (India) West of Longitude 72° 45′ E. G. DEP. COTTER. Pp. 161, pls. 9. India Geol. Surv. Mem. LV, Pt. 2. Calcutta, 1933. 8/6.

Pyu Earthquakes of 3d and 4th December, 1930, and Subsequent Burma Earthquakes up to January, 1932. J. C. BROWN AND P. LEICESTER. Pp. 140, pls. 6, figs. 3. India Geol. Surv. Mem. LXII, Pt. 1. Calcutta, 1933. 7/.

Vindhyan Sedimentation in the Son Valley, Mirzapur District. J. B. AUDEN. Pp. 110, pls. 10, figs. 9. India Geol. Surv. Mem. LXII, Pt. 2. Calcutta, 1933. 8/6.

Barytes in the Ceded Districts of the Madras Presidency: With Notes on its Occurrence in Other Parts of India. A. L. COULSON. Pp. 142, pls. 5, figs. 2. India Geol. Surv. Mem. LXIV, Pt. 1. Calcutta, 1933. 6/6.

Gold-Quartz Veins South of Libby, Montana. R. GIBSON. Pp. 25, figs. 3. U. S. Geol. Surv., Circ. 7. Washington, 1934.

SCIENTIFIC NOTES AND NEWS

Sydney H. Ball has been visiting Ontario and British Columbia on professional business.

John G. Barry has been appointed special investigator for the copper industry by Administrator Johnson of the N. R. A.

M. King Hubbert, of Columbia University, is in charge of a geophysical investigation for the U. S. Geological Survey in the Kentucky-Illinois fluorspar district.

Frank G. Stevens is general manager of Central Porcupine Mines Ltd., a company formed for exploration between the Dome and McIntyre Mines in Porcupine. D. G. H. Wright is acting as geologist.

W. H. Newhouse, of the Massachusetts Institute of Technology, is now the chairman of the Committee on Processes of Ore Deposition of the National Research Council.

Sir John Cadman, of the Anglo-Persian Oil Company, has been appointed President of the Institute of Fuel, London, and will assume office on October 17.

W. L. Whitehead, of the Massachusetts Institute of Technology, is inspecting properties of the Utah-Apex Mining Company on the Mother Lode, California, and at Bingham, Utah.

A. H. Koschmann, of the U. S. Geological Survey, is spending the summer in the Ketchikan district, Alaska, investigating for the Survey the known mineral occurrences of that region.

Geo. C. Branner was recently elected president of the Association of American State Geologists.

Peter Price has received from McGill University, Montreal, the Governor General's Medal.

T. L. Walker has been elected Vice-President of the Royal Society of Canada, and Cyril Knight and J. A. Allan have been elected President and Vice-President, respectively, of Section IV (Geological Sciences).

A Planning Committee for Mineral Policy was appointed by President Roosevelt in April, to undertake a general study of the mineral resources of the United States, with a view to the establishment of a national mineral policy. The committee will consider the estimating of future consumption, curtailment of production, co-ordination of emergency appropriations, relationship between Federal and State control, conservation of mineral resources, and the effect of mineral tariffs. The members

are: Harold L. Ickes, Secretary of the Interior, chairman; C. K. Leith, vice-chairman; Herbert Feis, Economic Adviser, Department of State; Lt. Col. C. T. Harris, Jr., War Office; Leon Henderson, Economic Adviser, N. I. R. A.; W. C. Mendenhall, Director, U. S. Geological Survey; Scott Turner, Director, U. S. Bureau of Mines; F. A. Silcox, Chief Forester; Wayne C. Taylor, Adviser, Foreign Trade; Willard L. Thorp, former Director, Bureau of Foreign and Domestic Commerce; J. W. Furness, Chief, Minerals Division, Department of Commerce.

The National Research Council has recently made the following grants in geology: Bradford C. Adams, Los Angeles, Calif., The Foraminifera of a Pliocene Section at Canada de Aliso, Ventura County, California; Gilbert H. Cady, Illinois Geological Survey, The Plant Components in Illinois Coal; George B. Cressey, Syracuse University, Geographic Field Work in China; Charles E. Decker, University of Oklahoma, Studies on Graptolites; V. C. Finch, University of Wisconsin, Preparation of a Series of Isopleth Maps of the United States; Harold L. Geis, University of Chicago, The Taxonomy of the Pennsylvania Ostracoda of Illinois; J. F. Lutz, No. Carolina College of Agriculture, Physical and Chemical Properties of Soils affecting Erosion; Evans B. Mayo, Cornell University, The Granites of the Eastern Sierra Nevadas; William F. Prouty, University of No. Carolina, The Silurian Deposits of Eastern Tennessee; Parry Reiche, University of California, The Lithology, Structure, and Sequence of the Sur Series of California at its Type Locality; Francis P. Shepard, University of Illinois, The Submarine Canyons off the California Coast; Samuel Weldman, University of Oklahoma, Dolomitization, Silicification, and Related Phases of Mineralization Associated with the Zinc-Lead Deposits of the Tri-State District; E. P. Wheeler, Ithaca, N. Y., A Study of the Anorthositic Rocks in the Vicinity of Nain, Labrador; George W. White, University of New Hampshire, Mapping of the Wisconsin-Illinois Glacial Boundary in North and West Central Ohio, by a Study of Soil Minerals.

The Council will consider further requests for research grants in the Fall. Application blanks will be furnished by the Secretary of the Committee on Grants-in-Aid, and should be filed with the Committee before October 15, 1934.

Jacob Johannes Sederholm, a geologist of note in Europe and also in this country, where he made a special study of the pre-Cambrian formations, died on June 27 at Helsingfors, Finland.

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